

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi

Zegeye Dressie

Recharge and Soil Water Studies Using Different Models and Measurement Methods

Uppsala University
Department of Physical Geography
Hydrological Division

Report Series A
Nos 2 and 39



Digitized by the Internet Archive
in 2026

https://archive.org/details/IA41551502_0006

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi

Zegeye Dressie

Recharge and Soil Water Studies Using Different Models and Measurement Methods

Uppsala University
Department of Physical Geography
Hydrological Division

Report Series A
Nos 2 and 39

Doctoral thesis at Uppsala University

ISSN 0281-8264

Copyright © 1987, by Zegeye Dressie

Printed in Sweden by

Reklam & Katalogtryck AB

Uppsala 1987

Dedicated to

My Beloved Wife and Children

Elisabeth, Johannes and Emanuel

Abstract

Dressie, Z. 1987. Recharge and soil water studies using different models and measurement methods. *Report Series A Nos 1984:2 and 1987:39*. 232 pp. Uppsala. ISSN 0281-8264

On a global scale, 98.5% of the water suitable for human consumption comes from ground water reservoirs which are refilled with water that percolates through the unsaturated zone. It is, therefore, important to understand the infiltration/percolation processes and to estimate the yearly recharge into the ground water storage.

In the first part of this thesis, percolation rates and ground water recharge were evaluated for three selected sites using the tritium-tagging and the water balance methods. Percolation rates of 2.1 to 2.3 mm/day during summer and 21.0 to 26.0 mm/day just after the snow melt were observed. Predicted areal variations of summer recharge values were between 13.2 mm and 79.4 mm.

The second part of the thesis deals with the application of a simulation model to study soil water flow and recharge estimates. Moisture contents as estimated by the model, the neutron probe, the TDR (Time Domain Reflectometry) and the gravimetric methods showed good agreement except near the surface of the soil and at very dry or wet conditions. The simulated, long term average yearly recharge values of 161 and 192 mm with the corresponding evapotranspiration losses of 452 and 352 mm for Skediga and Uppsala sites respectively, are in the same range of values as those predicted by other workers using the water balance method. A high correlation between the winter, autumn and annual recharge estimates and the corresponding precipitation values was established after regression analysis.

The simulation model proved to be very important when the temporal and the spatial patterns of soil water flow and recharge changes are of interest.

Zegeye Dressie, Division of Hydrology, Department of Physical Geography, Uppsala University, P.O. Box 554, S-751 22 Uppsala, Sweden.

ISSN 0281-8264

© Zegeye Dressie 1987

Printed in Sweden 1987
Reklam & Katalogtryck AB
Uppsala, Sweden.

P R E F A C E

The quaternary deposits, specially the glaciofluvial materials with sand and gravel as the dominant components (eskers), covering the major part of Sweden are extremely important for the formation and occurrence of ground water. Several previous attempts have been made to estimate recharge to ground water reservoirs using the water balance method.

The method of injected tritium profiles to follow the vertical movement of soil moisture in the unsaturated zone was started in the spring 1974 at a site 6 km east of Lycksele town ($64^{\circ} 35' N$, $18^{\circ} 40' E$) on the southern bank of the river Umeåälven. The idea was initiated by professor Erik Eriksson of the Division of Hydrology (at present Professor emeritus). With the experiences from Lycksele site, the method was further extended to other areas with similar formations (Skediga spring 1975, Jädraås spring 1976 and Tärnsjö spring 1977) in order to evaluate recharge estimates and study the vertical soil moisture movement.

During 1982 - 83, a sandy formation, in the vicinity of Uppsala town was selected and a combined method of tracers and a simulating model was used to study recharge and soil water flow. The model was provided by Dr Per Erik Jansson from the Swedish University of Agricultural Sciences, Department of Soil Sciences. The later part of the work was accomplished under the supervision of professor Lars Bengtsson of the Division of Hydrology. The cooperation of Kjerstin Andersson and Assar Lindberg with the drawing and the photographing work was very valuable. I also wish to express my full gratitude to all those, who, in some way or another, have rendered their support and encouragement.

The project was financially supported by the Swedish Natural Sciences Research Council (NFR) and the University of Uppsala.

This work could not have been accomplished without the support, understanding and patience of my wife.

PART I

UNIVERSITY OF UPPSALA

DEPARTMENT OF PHYSICAL GEOGRAPHY

DIVISION OF HYDROLOGY

ESTIMATION OF GROUNDWATER RECHARGE IN ESKER FORMATIONS USING THE ^3H -
TAGGING AND THE WATER BALANCE METHODS; STUDY OF SOIL MOISTURE MOVEMENT.

ZEGEYE DRESSIE

UPPSALA, 1984

	TABLE OF CONTENTS	PAGE
1	INTRODUCTION	1
2	GROUNDWATER RECHARGE	9
2.1	Different ways of groundwater recharge	10
2.1.1	Natural recharge	11
2.1.2	Artificial recharge	12
2.2	Models applied	13
2.2.1	The inventory method	13
2.2.2	The tritium-tagging method	14
3	THEORY OF SOIL MOISTURE FLOW	18
3.1	General: saturated/unsaturated	18
3.2	Unsaturated flow	19
3.3	Flow in layered soils	21
3.4	Non-steady-state flow	26
4	VARIABLES USED AND THEORY OF FIELD MEASUREMENTS	33
4.1	Precipitation	33
4.2	Evapotranspiration	34
4.3	Soil moisture determination	39
4.4	Tritium profiles	46
5	MEASUREMENT TECHNIQUES USED	48
5.1	Precipitation	48
5.2	Evapotranspiration	48
5.3	Soil moisture measurement	49
5.4	Soil profile studies	49
5.5	Tritium measurements in the field	50
5.5.1	Selection of sampling points	50
5.5.2	Method of injection	53
5.5.3	Injected tritium sampling method	55
5.5.4	Tritium analysis - Tritium laboratory	56

5.5.4.1	Injected tritium	56
5.5.4.2	Natural tritium	58
6	THE HYDROGEOLOGY OF THE INVESTIGATION AREAS	62
6.1	Jädraås	62
6.2	Skediga	62
6.3	Tärnsjö	65
7	MEASUREMENTS AND ANALYSIS	67
7.1	Soil profiles	67
7.1.1	Jädraås	67
7.1.2	Skediga	68
7.2	Evapotranspiration and precipitation data	70
7.2.1	Jädraås	70
7.2.2	Skediga	70
7.2.3	Tärnsjö	71
7.3	Soil moisture	74
7.3.1	Jädraås	74
7.3.2	Skediga	75
7.3.3	Tärnsjö	77
7.4	Tritium profiles	82
7.4.1	Jädraås	82
7.4.2	Skediga	84
7.4.3	Tärnsjö	88
8	RESULTS	96
9	DISCUSSION AND CONCLUSION	106
10	SUMMARY	110
11	APPENDIX	112
*	ACKNOWLEDGMENTS	133
*	REFERENCES	134

1. INTRODUCTION

The purpose of this investigation is to estimate groundwater recharge to esker formations by following the percolate water down the soil profile by using injected tritium as a tracer and when possible to compare the tritium-tagging method with the water balance method. The soil moisture content variations with time are monitored by using the neutron moderation method.

The use of groundwater for human consumption is a very old practice going back to antiquity. [17,18,63,71]. In the arid and semi-arid lands of the east, the near east and the Mediterranean basin, man learned to tap vast reservoirs of groundwater by digging wells, commonly several feet in diameter and by driving tunnels called kanats into the alluvial fans festooning mountain chains such as the Atlas and the Himalayas [17]. The Chinese learned the art of drilling wells to artesian wells that are reported to have been as deep as 5000 feet. The first drilled wells in Europe were found in Artois in France from which the name "artesian" originated and in Modena in northern Italy [63]. The use of groundwater has tremendously increased after 1900 due to increased demand of water for irrigation, industry and household which is accompanied by the rapidly rising standard of living. Furthermore improvements in construction of wells and pumps and the reduction in cost of power from electricity, gas and oil have all contributed to the extended exploitation of groundwater.

Out of the total amount of 1350 million Km^3 of water in the hydrosphere, only 8.2 million Km^3 is fit for human consumption as fresh water, and 98.5 percent of this amount comes from groundwater reservoirs [38]. At present it is an accepted fact that the origin of groundwater is mainly water from precipitation which has percolated down through the soil profile. Natural subsurface reservoirs are large and cheap, they are also less exposed to evaporation and pollution.

This increasing dependence on groundwater reservoirs for water supply calls for a thorough understanding of aquifer storage systems and recharge rates. Conventional methods such as the water balance or the inventory method, the lysimeter method and the storage method have been widely used in the estimation of groundwater recharge. For a well defined catchment area, if one has the precipitation, surface- and subsurface flow and the evapotranspiration records, recharge can easily be estimated using the water balance equation which can be expressed as:

$$R = P - (SR + O) - ET \quad (1)$$

In equation (1) R = recharge, P = precipitation, O and SR = subsurface and surface flows respectively, ET = potential evapotranspiration. Precipitation can be measured in the field, the surface and subsurface flows can be estimated. The potential evapotranspiration can be calculated by using any of the various mathematical formulas (for example the Penman equation) but an accurate value of the actual evapotranspiration is difficult to obtain. The lysimeter method does not completely represent the natural conditions. At the same time, since the recharge water passing through the soil column changes chemically and speedwise, it is essential to understand the processes in the unsaturated zone. None of the above mentioned methods takes account of this important process in the unsaturated zone.

The isotope tracer method is one way of directly estimating recharge by following the percolating water. Frequently used radio isotopes in hydrological studies are tritium (T , 3H), carbon-14, silicon-32 with half-lives of 12.3, 5730 and 500 years respectively. The possible use of atmospheric tritium in hydrological studies has been postulated and applied by several workers [19,20,21,22,31,39,78,79]. Isotopic methods of groundwater studies can be:

- (a) groundwater dating method
- (b) environmental tritium method
- (c) artificial tagging method

The groundwater dating method uses the relationship between the volume of groundwater and the residence time which can be obtained from the transit time distribution of water at several points in the aquifer.

The equation for this method is: $R = V / t \quad (2)$

In equation (2)

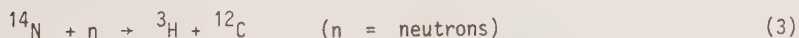
R = recharge rate

V = groundwater volume

t = the time the water spends in the reservoir

In the environmental tritium method the peak concentration of the tritium input in to the atmosphere by the nuclear bomb explosions which started in 1952 is traced in the lithosphere and the time of penetration to the observed depth is then used for the estimation of the percolation rate of moisture. The maximum concentration observed in 1963 in the northern hemisphere amounted to well over 1000 TU. (1 TU being equal to the ratio of 1 tritium atom to 10^{18} hydrogen atoms, ie. $1^3\text{H}/10^{18}.1\text{H}$). The bomb tritium in the atmosphere has been continuously decreasing so that at present it is approaching the natural concentrations.

Tritium is naturally produced by the cosmic ray action on nitrogen atoms according to the following reaction.



The pre-bomb natural concentration of tritium in the atmosphere was estimated to be 5 to 10 TU.

The vertical movement of soil moisture can be traced using injected tritium as a tracer. This method is based on the assumption that soil moisture moves vertically downwards in layers without bypassing the layer ahead. Such a flow is termed as "piston flow". This means that if a tritium peak is located at a depth z_2 below the injection depth z_1 , as much water as is between z_1 and z_2 is pushed down into groundwater zone. The equation describing the process can be written as:

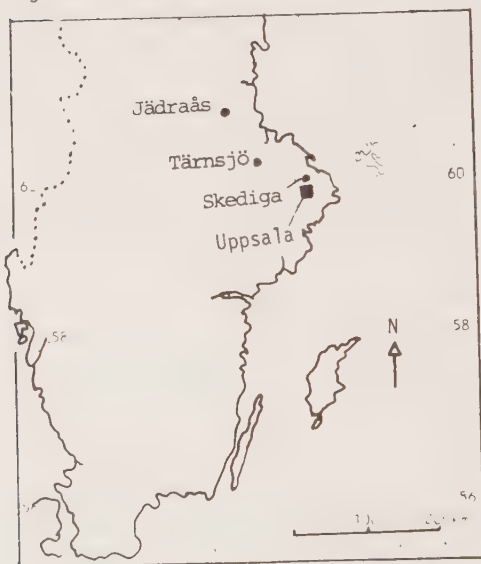
$$\bar{R} = (z_2 - z_1) \times \bar{\theta} \quad (4)$$

where z_1 and z_2 are the depths of the tritium peaks at times t_1 and t_2 , $\bar{\theta}$ is the soil moisture content of the profile between z_1 and z_2 at time t_2 and G is the groundwater recharge.

Equation (4) is used in the tritium-tagging method applied in this investigation. The method has been used and tested by several workers. [7,778, 79].

The location of the three areas of investigation is shown in fig. 1. They all lie in the extensive glaciofluvial formations of the Swedish soil system. With the exception of some outcrops, almost the whole of

Fig. 1 Location of the three investigation areas.



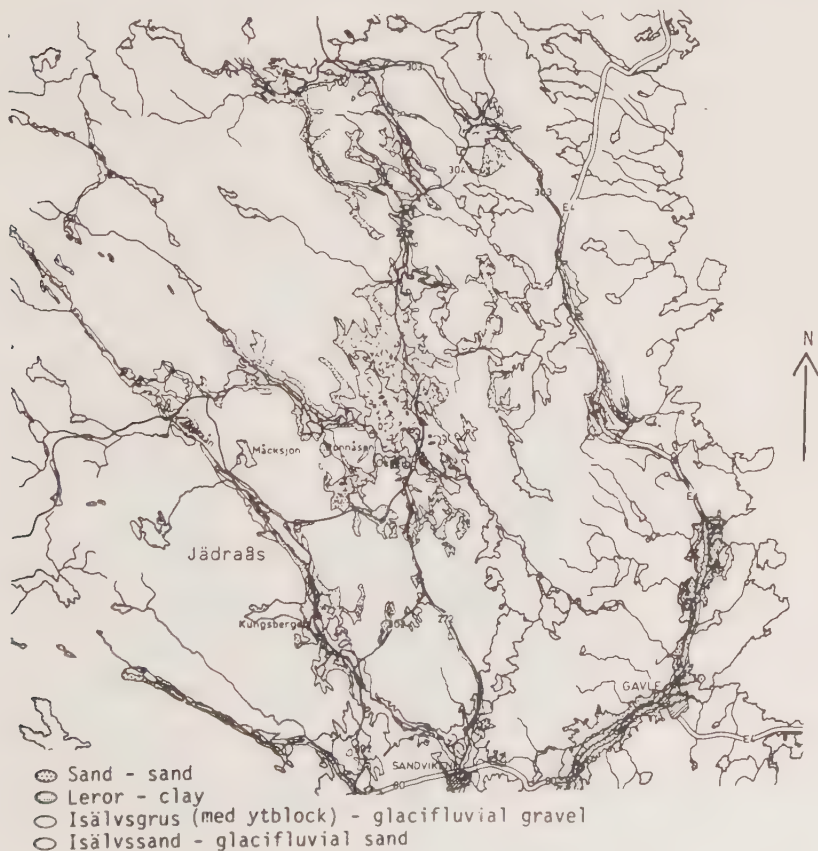
Sweden is covered with quaternary deposits (Moraines and eskers). Glaciofluvial deposits with sand and gravel as the dominant materials (eskers) make such formations which are so important for the formation and occurrence of groundwater. For example more than 70% (ie. 500 l/s out of 710 l/s in 1971) of the water supply of Uppsala town comes from Uppsala esker. [60].

The method of injected tritium was started in Lycksele in the spring of 1974. The station sel-

ected had a homogeneous sandy formation and fairly good tritium profiles were obtained. Under the auspice of Prof. Erik Eriksson of the Division of Hydrology, University of Uppsala, and with the experience from the Lycksele experiment, it was planned to extend the method to other areas of glaciofluvial formations. Consequently, starting from the summer of 1976, tritium injections were done in the two areas Jädraås and Skediga with Tärnsjö coming into the picture in the summer of 1977.

Jädraås: This area of research is located near Jädraås at Invartjärnsheden, in the narrow valley of river Jädraån and to the west of it. See fig. 2. Invartjärnsheden is part of the larger esker formation known as Järboåsen which is about 75 km long and stretches from Sandviken to lake Mällängen (outside the map, fig. 2) in SE to NW direction. The glaciofluvial deposit is underlain by the leptite, porphyry and gneiss-granite bedrock. [37].

Fig. 2 Map over the area around Jädraås with marine and glaciofluvial sediments. Scale 1:5000 000. (From Per Erik Jansson, 1977)



Skediga: This location lies on a system of a long ridge of glaciofluvial deposits known as eskers. The formation of interest is called Uppsala esker. The part of the formation which has been subjected to town water exploitation comprises of the following sections. See fig 3.

(a) South of Uppsala

1) Sunnerstaåsen - (Flottsund - Sunnersta area)

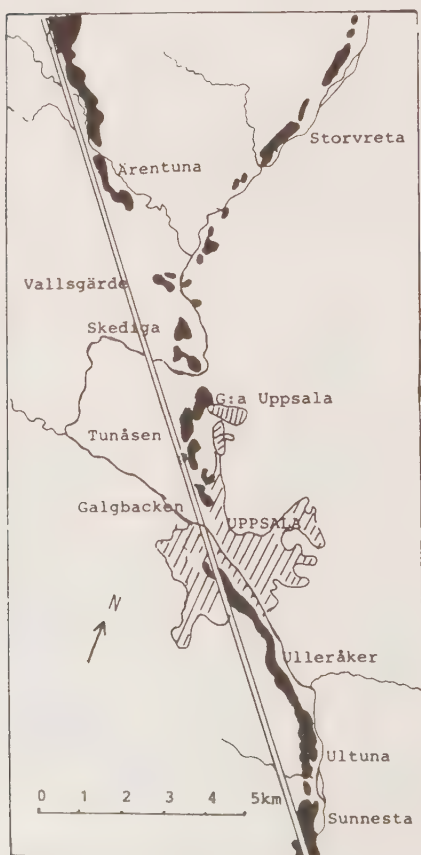
2) The long continuous (5 km) formation which comprises of the Ultuna area, the Ulleråker area, Polackbacken, Kronåsen and ends at Domkyrka before it is buried under the clay layer.

(b) North of Uppsala

3) Uppsala - Gamla Uppsala area comprising of Galgbacken, Röbo-Tegelbruk area, Tunåsen and Gamla Uppsala area.

4) Another small group of scattered esker formation where the investigation area (Skediga) is situated.

Fig.3 Uppsala esker in the vicinity of Uppsala town.



The esker formation of Uppsala shows distinct differences both in form and direction of formation. Some are scattered and do not follow the fault lines. The highest and well developed formation is the Tunåsen (50 m high and 700 m long). The esker at Skediga, where the experiment is run, is somehow elongated with a steep slope in the eastern edge and a very gradually rising western side.

Tärnsjö: The Tärnsjö area is situated on Enköpings esker (Enköpingsåsen) about 60 km north-west of Uppsala. This esker formation belongs to the series of ridges of the Eastern Swedish Esker formations. The locations of the different plots (A - E) with the elevations of the area, which ranges from 55 to 85 m above sea level are shown in the map fig. 4 taken from Norberg and Persson (1974).

Fig. 4 Map showing the location of the sampling points in Tärnsjö area. (Norberg & Persson, 1974)



The results of the experiment show that in heterogeneous materials such as eskers, the assumed vertical movement of percolating water is not always fulfilled because of the frequently occurring layered formations which can cause possible lateral flow. This condition of coarse-over-fine layering is distinctly seen in Skediga area. Hence a careful selection of a homogenous profile is essential for the application of tracer-tagging method. It is also found that the traditional sampling method with the use of porous cups can not be successfully applied in esker and moraine materials because of the poor contact between the porous cups and the soil material and due to the difficulties frequently encountered when installing the sampling tubes with the porous cups on. This led to the development of a new sampling method in which soil air is sucked and the vapour part is absorbed by molecular sieve-beds [16, 66].

The groundwater recharge estimated by the tritium-tagging method is compared with that estimated by the water balance method. Because of the lack of soil moisture data for all the stations in Tärnsjö area, the comparison of the two methods is done only for one plot, plot C. The soil moisture data from one common point is used in the water balance estimation of recharge for the two plots in Jädraås area. The two stations lie very close to one another. In spite of the heterogeneity of the formations and the difficulties encountered, the two methods give values which are in an acceptable range for most of the cases. However, a model which takes into account the dispersion/diffusion processes based on data for the whole year can certainly give a better result. With the necessary improvements of the model and the sampling method, tracer application is indispensable to study the complicated and poorly understood process of unsaturated moisture flow in the soil.

The possibility of insitu monitoring of the temporal variations of the soil moisture content using the neutron moderation method is also exploited in this experiment.

Both injected and environmental tritium methods have been widely used in hydrological studies by many workers and in many countries. To mention only a few of them, Anderssen (1974) in Denmark, Zimmermann (1967) in Germany and Athavale et al (1978) in India have done an intensive work in this field.

2. GROUNDWATER RECHARGE

Groundwater is one part of the earth's water cycle or the hydrologic cycle which is a continuous circulation of water in the form of liquid or vapour in our planet [56,68]. Since the oceans cover one quarter of the earth's surface, it is logical to begin the water cycle with the waters in the oceans. Referring to figure 5 which is a sketch of the hydrologic cycle, solar radiation evaporates water from the ocean surfaces and other water bodies (lakes and streams), the earth's surface and from the leaves of plants into the atmosphere. The evaporated moisture rises up, condenses and falls back to the surface of the earth as precipitation. Out of the estimated of 95 000 cubic miles of water evaporated, 80 000 cubic miles is from the ocean surfaces and 15 000 cubic miles is from the surface of the earth. [38]. As much water as total evaporation falls back as rain each year. Some of the precipitation that has fallen back reevaporates to the atmosphere, some of which contributes to surface runoff and the rest infiltrates into the soil. Much of the infiltrated water is detained in the soil root zone, drawn back and lost as transpiration. The rest, if any, percolates through the areated zone into the zone of saturation some of which may find its way to the surface of the earth or into other water bodies as discharge. That part of water that is fianally pulled down by gravity into the zone of saturation is known as recharge.

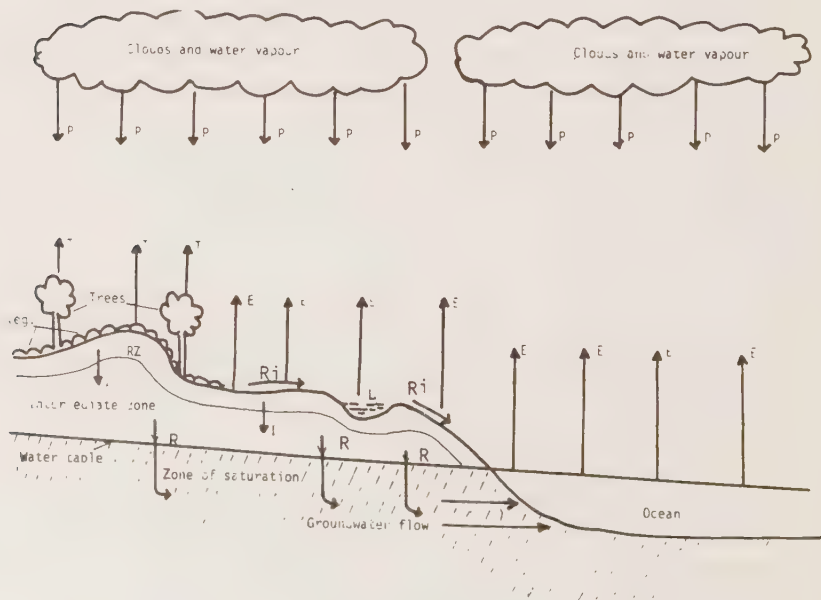
Water in the zone of saturation is commonly known as "groundwater" but terms such as "underground water" and "subterranean water" are also met. If a geologic formation contains pores which are filled with water and if the pores are big enough to allow water to move towards wells and springs at a perceptible rate, then the formation is called aquifer. (sometimes also called "water bearing formations" or "groundwater reservoirs") [68]. When the upper boundary of an aquifer is the groundwater level, then the aquifer is called water table aquifer. Some othe terms used are "unconfined aquifer", "unconfined groundwater aquifer" or "free groundwater aquifer". [38].

When an aquifer is bounded both below and above, by impermeable layers, it is called confined, artesian or confined groundwater aquifer. If a well is drilled through the upper confining layer, the water rises in the well to some level above the top of the aquifer and this level is called the piezometric level.

2.1 DIFFERENT WAYS OF GROUNDWATER RECHARGE

Groundwater can be replenished through the hydrologic cycle (precipitation), through contact with surface water and through ways induced by man (artificially).

Fig. 5 The hydrological cycle. Legend: P = precipitation, T = transpiration, E = evaporation, I = infiltration, RZ = root zone, L = lakes, Ri = rivers, R = ground water recharge



2.1.1 NATURAL RECHARGE

Natural recharge can occur from precipitation or through contact with surface water.

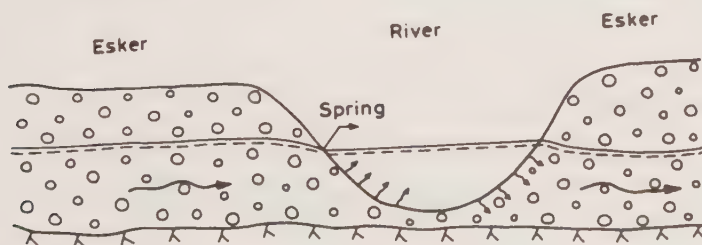
a) As we have discussed earlier, precipitation falling on the surface of the earth is dissipated in several ways. Part of the infiltrated water from precipitation, if not lost by evapotranspiration, percolates through the intermediate aerated zone into the saturated or groundwater zone as recharge. How much water infiltrates and how much becomes surface runoff depends upon the topography, inclination, soil material and the intensity of rainfall. In Swedish conditions recharge mainly occurs immediately after the snow melt. During the late summer months evapotranspiration loss is usually greater than precipitation. In the winter period precipitation is in the form of snow and the top layer of the soil is frozen and under saturated conditions, is impermeable for any infiltration to take place.

b) Recharge through stream beds.

When a stream cuts a sandy formation as shown in figure 6, infiltration into the underlying formation can take place. In this case the aquifer must be unconfined.

Fig. 6 Stream cutting water bearing formation.

(Taken from Gustafsson, 1977)



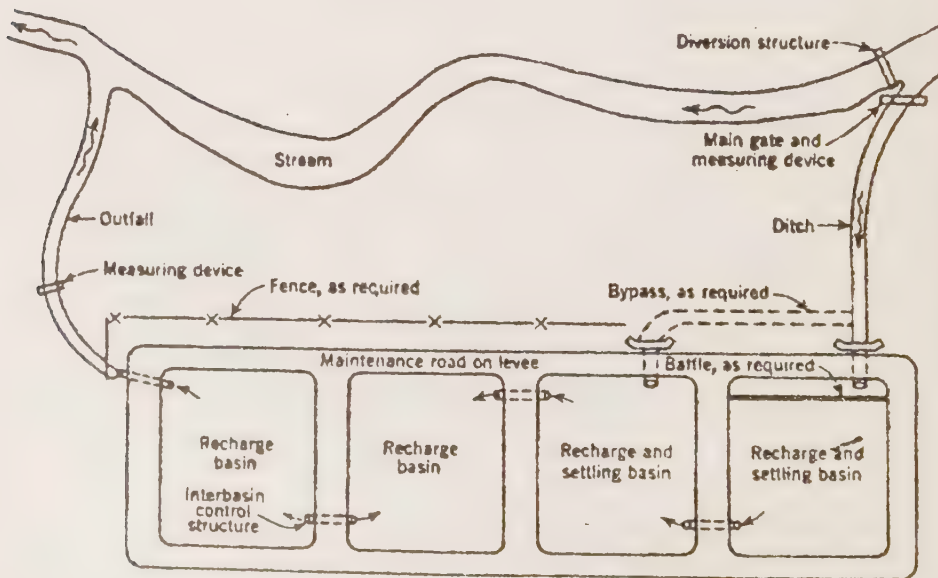
2.1.2 ARTIFICIAL RECHARGE

Due to the increasing exploitation of groundwater resources, the depletion of groundwater is going faster than the replenishment by natural means. Therefore, more and more work is being done to recharge groundwater artificially. Some of these artificial methods are listed below.

a) Recharge basins.

In this method, surface water is either pumped or diverted to specially designed sand beds through which water is allowed to infiltrate into the unconfined aquifer. One typical example is the water well gallery at Tunåsen about 2 km north of Uppsala town (Sidenvall, 1970) where water from Fyris river is pumped up and infiltrated into the esker formation. Fig. 7 shows a scheme where a stream is diverted to recharge basins. (Taken from Henriksson and Johnsson, 1971).

Fig. 7 A scheme for an artificial recharge from a stream



b) Recharge through wells.

This method is specially used for confined aquifers. Such recharge wells can also be used as pumping wells and are extensively used in Israel.

c) Recharge induced by pumping.

When water is pumped from an aquifer, the water level in the vicinity of the well falls making a drawdown line which forms a cone shaped depression round the well. This cone of depression extends away from the well as pumping continues and its shape and size depend upon the pumping rate, length of pumping time, aquifer characteristics, slope of the water table and the recharge within the zone of influence. When the radius of influence crosses a water body such as a lake or a river, water is drawn towards the pumping site, thus inducing recharge to the aquifer. For example both Holland and Israel have had problems of salt water intrusions induced by pumping near the coast.

2.2 MODELS APPLIED

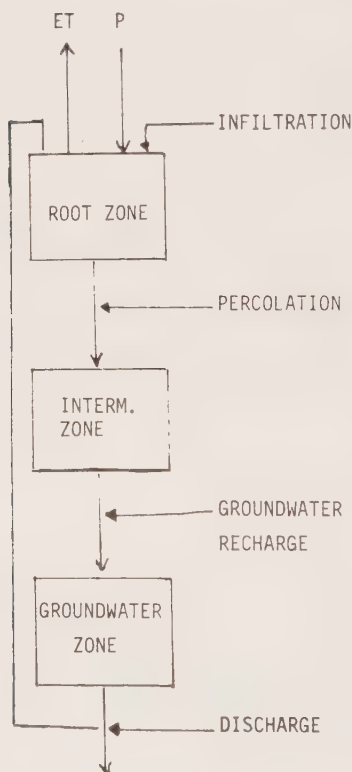
"A model, as a simplified concept of nature, is a tool one of the main purpose of which is to overcome difficulties in comprehending the complexity of natural processes". (Eriksson, 1975). One of such complex natural processes is groundwater formation by recharge processes, primarily because of the complicated soil moisture flow process which takes place as water passes through the unsaturated zone before it reaches the groundwater regime. The two models tested in this investigation are the inventory (or the water balance) and the tritium-tagging methods.

2.2.1 THE INVENTORY METHOD

The inventory method is based on the simple water balance equation which can be expressed as:

$$P = ET + SR + dS + R \quad (4b)$$

Fig. 8 A simple block diagram for the process of recharge formation. (Eriksson, 1975)



where,

ET = Evapotranspiration

dS = Soil moisture storage change

R = Groundwater recharge

SR = Surface runoff

P = Precipitation

Fig. 8 is a simplified conceptual model which can be described by the simple water balance equation. As indicated by the arrows, part of the precipitation (P) falling on a catchment area is lost by evaporation from the ground surface and by transpiration from plants and vegetation cover. The combined process is called evapotranspiration (ET). The rest of the precipitation after surface runoff (SR), if any, infiltrates into the root zone. Some of the infiltrated moisture may be taken up by plant roots and be lost by transpiration and the rest of it percolates through the intermediate zone into the groundwater zone as

recharge. Surface runoff (if any) and outflow from the groundwater

storage make up the discharge from the catchment area. If we rewrite equation (4b) in terms of recharge, we get,

$$R = P - (ET + SR + dS) \quad (4c)$$

From this relation we see that the water balance method is based on the principle that recharge is a difference between an input into the system

and an output from the system. The input part of the equation consists of precipitation and/or irrigation. In the present experiment the input is only precipitation. The output part is made up of all the other terms on the right hand side of the expression. In a very coarse material such as sandy formation, where the main flow direction is vertically downwards and no surface runoff occurs, equation (4c) reduces to:

$$R = P - (ET + dS) \quad (4d)$$

This equation stands for the inventory method applied in the present investigation.

2.2.2 THE TRITIUM-TAGGING METHOD:

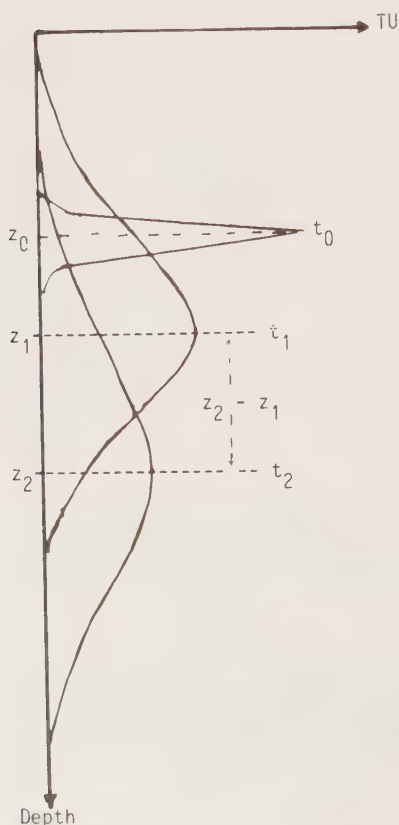
There are several types of tracers that can be used in hydrological studies. Besides the type and purpose of the investigation, there are other criteria to be considered when choosing the kind of tracer to be used. Such criteria and some of the different types of tracers are presented in the appendix. Concerning the choice of tritium, the most important points are that ^3H is part of the water molecule and moves with it with no retardation, that it has little radiation hazard, that it has a convenient half-life of 12.3 years and that it is easy to apply. Tritium can also be detected in very low concentrations using modern scintillation spectrometers. That tritium, as part of the water molecule moves with it without absorption or retardation by the porous media through which the water moves, is true for coarse materials such as glaciofluvial deposits (eskers) or moraines.

In clay soils containing montmorillonite and kaolinite some absorption and exchange of the tritiated water may take place between the clay crystals. (Corey, 1973).

It is also assumed that soil moisture movement is based on the theory of piston flow (Zimmermann et al, 1967). The theory postulates that soil moisture moves in layers without mixing with the preceding layer of moisture but by pushing it ahead like in a piston. In practice it means that moisture from the input of the previous precipitation arrives

at the point of interest without mixing with the moisture that already existed and before the arrival of the moisture from a later precipitation event. This in turn leads to the basic assumption of the method that water can not bypass or shortcut the tritium tagged layer in either direction. Therefore the tritium peak can be considered as a surface and since it is itself moving, it acts as a moving piston. Hence "piston flow". This makes it possible to write a water balance for the soil water column above the tagged layer. The displacement of the peak during a certain time interval gives us the rate of the soil moisture movement. Fig. 9 is a hypothetical diagram showing two tritium profiles

Fig. 9 Hypothetical ^3H profiles.



for the sampling occasions t_1 and t_2 with the corresponding peaks at the depths z_1 and z_2 . If the average rate of the vertical movement of soil moisture as indicated by the tritium peak displacement is denoted by $\bar{R}$, then the following relation is valid.

$$\bar{R} = \frac{\int_{z_1}^{z_2} \theta \, dz}{t_2 - t_1} \quad (4e)$$

where,

$\bar{R}$ = rate of moisture movement
in $\text{m}^3/\text{s} \cdot \text{m}^2$. (i.e. recharge)

θ = soil moisture content by
volume fraction at time t_2 .

The position of the tritium peak marks the depth to which soil moisture has moved in the vertical direction.

If the displacement of the peak is $z (= z_2 - z_1)$ cm and the average soil moisture content by volume fraction is $\bar{\theta}$, then the amount of moisture added to the profile is equal to the product of the displacement of the peak and the average soil moisture content. Applying the principles of piston flow an equal amount of moisture is added to the groundwater storage. Hence the equation for the model can be written as,

$$R = z \cdot \bar{\theta} \quad (4f)$$

If the unit of the displacement is in mm, then the recharge is also in mm in other words the added recharge has the same unit as the displacement (z) of the peak.

3. THEORY OF SOIL MOISTURE FLOW

3.1 GENERAL: SATURATED/UNSATURATED

In order to understand the behaviour of the tritium profiles, it is essential to take up some of the problems and principles connected with the processes of soil moisture movement. In saturated flow through a porous media, the flux is caused by a potential gradient and is affected by the geometric properties of the pore channels. Darcy's law for the saturated moisture flow reads,

$$q = -K \frac{dH}{dz} \quad \text{and} \quad H = (\psi + Z) \quad (5)$$

Where q is the flux of moisture, K is the hydraulic conductivity, H is the total hydraulic pressure or hydraulic head, ψ is the pressure head, and Z is the height from a reference level. Hence it is clear that the driving force in a saturated flow is a positive pressure gradient. It is somewhat different with unsaturated flow. The same principle holds here too but the driving force, instead of being positive pressure gradient, it is a negative suction force, also known as matric suction or capillary suction and is mainly a sum of the adhesion and capillary forces.

Another important feature of soil moisture flow is that in unsaturated flow the conductivity changes with the change in soil moisture content whereas in saturated flow the conductivity has its maximum value. The soil texture is also important in soil water movement. In coarse textured materials the bigger pores empty faster and unsaturated water movement occurs only in smaller pores as a result of which the conductivity values decrease. Hence, coarser materials hamper flow under unsaturated conditions and increase conductivity under saturated conditions. This also means that conductivity decreases with decreasing soil moisture content. (Phyllips, 1958, 1969; Richards, 1953; Gardner, 1958; Hillel, 1980; and Poullovassilis, 1969, 1974)

In general unsaturated soil moisture flow is a very complicated process mainly because of the interdependence of the soil moisture content (θ), the hydraulic conductivity (K) and the matric suction (ψ). The interrelationship among these variables are non-linear.

It is also clearly understood that although the primary forces causing soil moisture flow are gravity and/or capillary forces, other factors affecting soil moisture movement are temperature gradient, precipitation, evapotranspiration, irrigation, groundwater level and degree of saturation. Unsaturated flow can be in any direction. Thermodynamical forces (e.g. soil frost in winter) and evapotranspiration can cause an upward movement.

Unsaturated conductivity, $K(\theta)$, is not directly proportional to the moisture content (θ). The former decreases faster than the later. Several attempts have been made to find mathematical relationships between conductivity and soil moisture content (or matric suction). For example according to Richards and Weeks (1953),

$$\theta = 0.65\psi^{-0.30} \quad (\text{for specific soil}) \quad (6)$$

$$\psi = 660K^{-0.38} \quad \text{or} \quad \log \psi = \log 660 - 0.38 \log K \quad (7)$$

Gardner (1958) found the following relationship,

$$\log K = c \log \theta \quad (8)$$

Observe that all the relationships are curvilinear.

3.2 UNSATURATED FLOW

As indicated above, the classical equation for fluid flow in porous media can not be applied for unsaturated conditions and Darcy's law must be modified. Since the water content of soils changes with time, one has to combine Darcy's law with the continuity equation. Darcy's equation in vector form is written as:

$$\vec{q} = Q/A = -K\nabla H \quad (9)$$

Where,

$$\nabla = \left(\frac{\delta}{\delta x}, \frac{\delta}{\delta y}, \frac{\delta}{\delta z} \right) \quad (10)$$

H = total hydraulic pressure = $(\psi + z)$

For vertical flow equation (9) reduces to,

$$q = -K \frac{\delta}{\delta z}(\psi + z) = -K \frac{\delta \psi}{\delta z} - K \quad (11)$$

According to continuity the rate of change of soil moisture content is the same as the flux during a given time interval (assuming no evapotranspiration loss from the part of the soil column in question). The mathematical expression is:

$$\frac{\delta \theta}{\delta t} = - \frac{\delta q}{\delta z} \quad (12)$$

If we now combine Darcy's equation and the equation of continuity, we get the expression:

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta z} \left(K \frac{\delta \psi}{\delta z} \right) + \frac{\delta K}{\delta z} \quad (13)$$

As stated above, in unsaturated flow, the hydraulic conductivity K depends strongly on the moisture content or the matric suction of the soil. By the chain rule,

$$\frac{\delta \psi}{\delta z} = \frac{d\psi}{d\theta} \cdot \frac{\delta \theta}{\delta z} = \frac{1}{C_\theta} \cdot \frac{\delta \theta}{\delta z} \quad (14)$$

The character $C_\theta (= \delta\theta/\delta\psi)$ is called the specific water capacity. Combining equations (13) and (14) we get the following expression.

$$\frac{\delta}{\delta z} \left(D \frac{\delta \theta}{\delta z} \right) + \frac{\delta K}{\delta z} = \frac{\delta \theta}{\delta t} \quad (15)$$

In equation (15) $D (= K/C_\theta)$ is the hydraulic diffusivity or the soil water diffusivity. It is very difficult to give any general mathematical solution for the above given soil moisture equation (Hillel, 1980). Consequently most of the attempts to understand soil moisture flow problems have been focused on the study of the soil moisture characteristics, ie. the functional relation of the hydraulic conductivity (K) to the soil moisture content (θ) and the matric potential (ψ).

3.3 FLOW IN LAYERED SOILS

In homogenous porous media, such as a soil profile, where water moves vertically downwards at a constant flow rate, a steady-state moisture and potential profile is established. The general shape of the profile is characterised by the presence of an upper part of uniform moisture and potential profile and a lower part with a decreasing potential and increasing moisture content as the groundwater level is approached, Childs (1945). In layered soils the shape of the profiles differs depending upon:

- (a) the sequence of the layers
- (b) the rate of flow
- (c) the thickness of the layer and
- (d) the distance of the junction from the groundwater level.

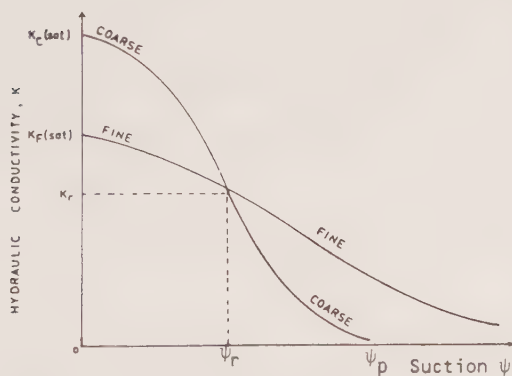
Generally, in layered soil profiles alternating uniform zones and zones with steep suction and moisture gradients may exist, Poulouvassilis (1980).

One way of experimentally determining the soil moisture characteristics of non-uniform soils is to establish a steady state flow by inducing a constant infiltration rate and looking at the resulting steady-state soil moisture and potential profiles, Poulouvassilis (1980). The non-uniformity of the soil is reflected in the non-uniformities of the suction and moisture gradients.

For the transient conditions, the moisture and suction gradients in a

drainig profile are studied, Poulovassilis (1969) and Ghuman et al (1980). The relative changes of the conductivity and suction in a two layered soil profile are shown by the hypothetical diagram below, Fig. 10

Fig. 10 A graph of hydraulic conductivity vesus suction for a layered soil of coarse and fine material.



Hypothetical relationships between the hydraulic conductivity and suction of two porous materials.

From the figure it is seen that the saturation hydraulic conductivity (K_s) is higher in the coarser material than in the finer one and it is also drained faster in the former than in the latter. Hence the steeper curve for the coarse material. When the two curves cross each other at (K_r, ψ_r) , the suction and hydraulic conductivities are equal. At higher suction values than ψ_r the finer material has a higher conductivity values. After a further lapse of time and at still higher suction values the hydraulic conductivity of the coarser material approaches the value of zero. The suction ψ_p in the figure corresponds to the zero K value for the coarse material.

According to Poullovassilis (1980), using figure 10 and the flow equations at the junction of a two layered profile, one can construct suction and soil moisture profiles for the soil in question. If we assume a constant flow rate, q , then the fluxes on both sides of the junction are equal. If $F(a)$ and $F(b)$ are the fluxes just above and just below the junction respectively, then

$$F(a) = F(b) = K_1(a) \cdot (dH/dz)_a = K_2(b) \cdot (dH/dz)_b \quad (16)$$

Where,

$K_1(a)$ and $K_2(b)$ = hydraulic conductivities in layer 1 and layer 2 respectively

$(dH/dz)_a$ = potential gradient just above the junction

$(dH/dz)_b$ = potential gradient just below the junction

If we now replace the hydraulic potential H by $(\psi+z)$ and also assume that the thickness of the lower layer is big enough, then the following equations can be derived.

$$d\psi/dz)_a = 1 - \{K_2(b) / K_1(a)\} \quad (17)$$

If we now assume a sequence of layering with a coarse material at the top of a finer material and by labeling the coarse and fine materials 1 and 2 respectively, the following relationships between suction ψ , hydraulic conductivities of the two layers K and the constant flow rates q can be deduced from fig.10

$$\psi > \psi_r \quad K_1(a) < K_2(b) \quad \text{and} \quad q < K_r \quad \psi_1 < \psi_2 \quad (18)$$

$$\psi = \psi_r \quad K_1(a) = K_2(b) \quad q = K_r \quad \psi_1 = \psi_2 \quad (19)$$

$$\psi < \psi_r \quad K_1(a) > K_2(b) \quad q > K_r \quad \psi_1 > \psi_2 \quad (20)$$

Poulovassillis used equation (17) and the relations in the equations (18), (19) and (20) to construct the hypothetical potential and moisture profiles for the different layerings. See figs. 11 and 12. For instance when $\psi_i > \psi_r$ and $q < K_r$ then, from equation (18) $K_1(a) < K_2(b)$ and applying this relationship in equation (17) gives $(d\psi/dz) < 0$. This implies that the potential increases when z approaches the junction. Hence curve ABC in fig. 11(a). Using the same reasoning curves DEF, GHL etc are constructed for $\psi_i = \psi_r$, $\psi_i < \psi_r$ respectively. Curves MNR and PQR are for $q = K_2(\text{sat})$ and for $K_1(\text{sat}) > q > K_2(\text{sat})$. Fig. 11(b) shows the corresponding moisture profiles for the same profile. Similar curves can be constructed for a fine-over-coarse layering sequence, Poulovassillis (1980). Hence if a steady state suction profile is established in a soil profile a clear information about the layering of a soil profile can be obtained. Fig. 12 is constructed for fine-over-coarse layered soil column.

Fig. 11 Hypothetical suction and moisture profiles for a steady state and coarse-over-fine layered soil column. (Adopted from Poulovassilis, (1980).

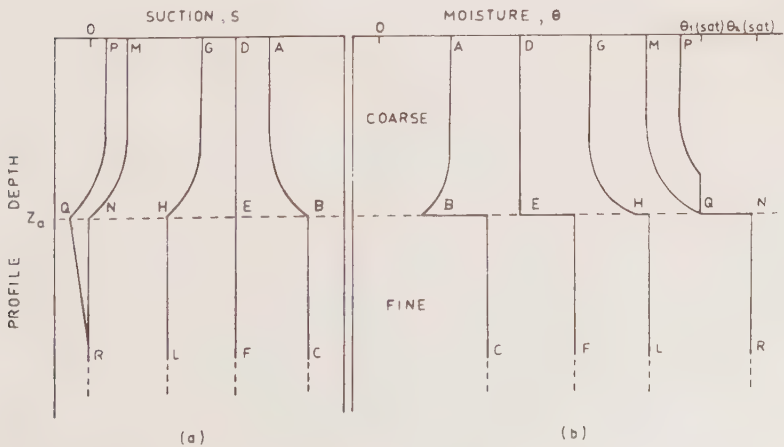
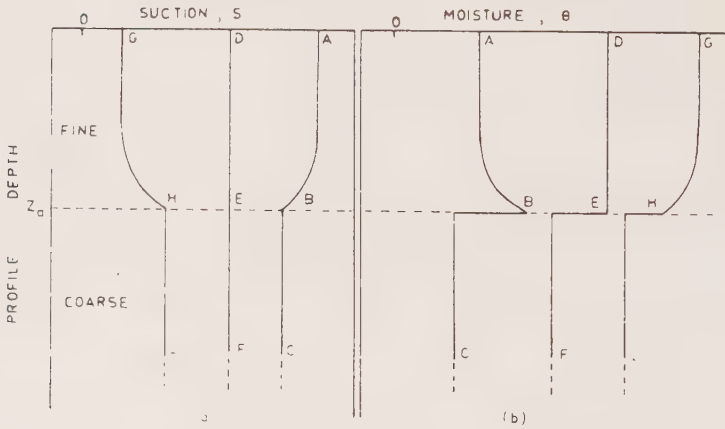


Fig. 12 Hypothetical suction and moisture profiles for a steady state and a fine-over-coarse layered soil column. (from Poulouvassilis, 1974)



3.4 NON-STEADY STATE FLOW

The steady-state flow method considered above can give a good picture about the soil moisture characteristics and consequently about the soil physical properties such as layering. In natural conditions soil profiles are usually in an unsteady and unsaturated conditions. Therefore flow processes have to be studied by the unsteady-state profile method, Ghuman and Prihar (1980) and Poulouvassilis (1980). In this method the flux is considered to be solely vertical and downwards. The evaporation loss at the surface is prevented by using plastic cover on the surface. Time t after the occurrence of infiltration and at the depth z below the surface, the flux of the soil moisture is considered. Under such conditions the flux F at the selected depth is given by the relation,

$$F = \int_z^0 (d\theta/dt) dz \quad (21)$$

According to Darcy,

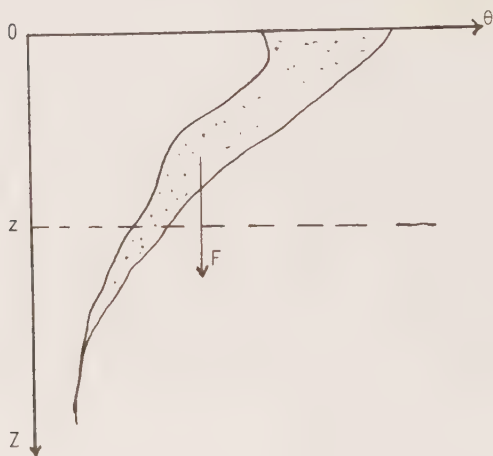
$$F = -K (dH/dz)_z \quad (22)$$

Combining the above two equations and solving for K gives.

$$K = -\left\{ \int_z^0 (d\theta/dt) dz \right\} / (dH/dz)_x \quad (23)$$

Thus from equation (23) we can see that if one knows the rate of soil moisture change with time and depth and the potential changes of the profile, the hydraulic conductivity K can be determined. The flux part of the equation (the nominator in the right hand side) can be estimated in terms of integrated fluxes or in terms of instantaneous fluxes. In the former method, the difference in moisture content at all depths as shown by the two moisture profiles for t_1 and t_2 is calculated. (fig. 13). (Rose et al, 1965). In the instantaneous flux method, the soil profile is divided into smaller slices of thickness Δz and $d\theta/dt$ for each slice and time t is determined. The rate of flux for each slice is then $(d\theta/dt)\Delta z$.

Fig. 13 Soil moisture profiles for two measuring occasions.



The total flux during the time t is then the sum of all the contributions of the individual slices and is expressed by the the equation below

$$F = \sum_z^0 (d\theta/dt)\Delta z \quad (24)$$

If we now assume a soil profile with a fine-over-coarse layering with a junction L as shown in fig. 14, the flux at the junction is given by the following relation. (Hillel, 1980).

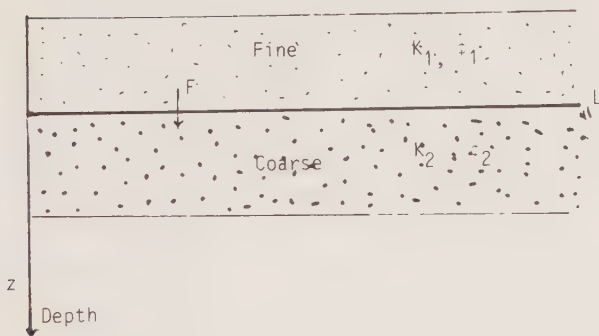
$$F_L = \int_{Z_L}^0 -K(dH/dZ) \quad (25)$$

The flux just above the junction, $F(a)$, is equal to the flux just below the junction $F(b)$ and can expressed as

$$F(a) = -(K_1(a)(dH/dz))_a \quad (26)$$

$$F(b) = -(K_2(b)(dH/dz))_b \quad (27)$$

Fig. 14 A fine-over-coarse layered soil profile



In the same manner as in the steady-state flow method discussed above, we can assume that the flux on both sides of the junction is the same, ie. $F(a) = F(b)$. Therefore,

$$K_1(a)(dH/dz)_a = K_2(b)(dH/dz)_b \quad (28)$$

And after rearranging we get,

$$K_1(a)/K_2(b) = (dH/dz)_b/(dH/dz)_a \quad (29)$$

In the case of unsteady-state, flow rate q is not constant and the flux will continuously decrease whereby the suction value increases.

In a fine-over-coarse layered formations where the lower layer is sufficiently thick, due to the lower rate of decrease of flux in the finer layer this change of flux may not affect the potential in the coarser material for a short time. After further increase of time, the suction value approaches ψ_p and at this point the potential gradients on both sides of the junction are equal (because $K_1(a) = K_2(b)$ at this point). At higher values of suction $K_1(a)$ becomes increasingly larger than $K_2(b)$ (See fig.10). To compensate for this change on the left hand side of equation 29, the potential gradient $(dH/dz)_b$ must increase and $(dH/dz)_a$ becomes smaller. Consequently the suction in layer 1 (finer material) approaches the equilibrium profile at the junction.

Because of this higher tension in the upper layer no moisture passes into the lower layer of coarser material until enough moisture has accumulated to decrease the suction value at the junction to a value lower than that in the coarser material. When the hydraulic conductivity K_2 of the lower layer (coarser material) becomes zero at suction ψ_p (see fig.10), the suction at the surface becomes,

$$\psi_s = \psi_p + X_1 \quad (30)$$

Where,

ψ_s = suction at the surface ($z = 0$)

ψ_p = suction when K_2 is equal to zero

X_1 = thickness of the upper finer layer.

Also, the coarser the second layer, the smaller is the ψ_p value. Consequently the total volume of water draining from layer 1 depends upon the thickness of layer 1 and the degree of coarseness of layer 2.

These theoretical considerations of flow in layered soils have been experimentally proved by several workers. Poulouvassilis (1980) has applied both the steady-state and the transient methods to study the flow of moisture in layered soils. The profile he studied is a coarse-fine-coarse formation (fig. 15 A). The suction and moisture profiles of his investigation are shown in figs. 15B - 15D. Both the steady-state (fig. 15C) and the transient (fig. 15D) suction profiles clearly reflect the layering sequence of the soil formation.

Ghuman and Pihar (1980) have also done an investigation of flow processes both in fine-over-coarse and coarse-over-fine layered formations. Of special interest is the effect of layering on tracer displacement. By using chloride as a tracer in the different formations they found the tracer movement in the vertical profile to be affected by the different formations in different manners. (see figs. 16A & 16B). From these investigation results and theoretical discussions, the following conclusions can be drawn.

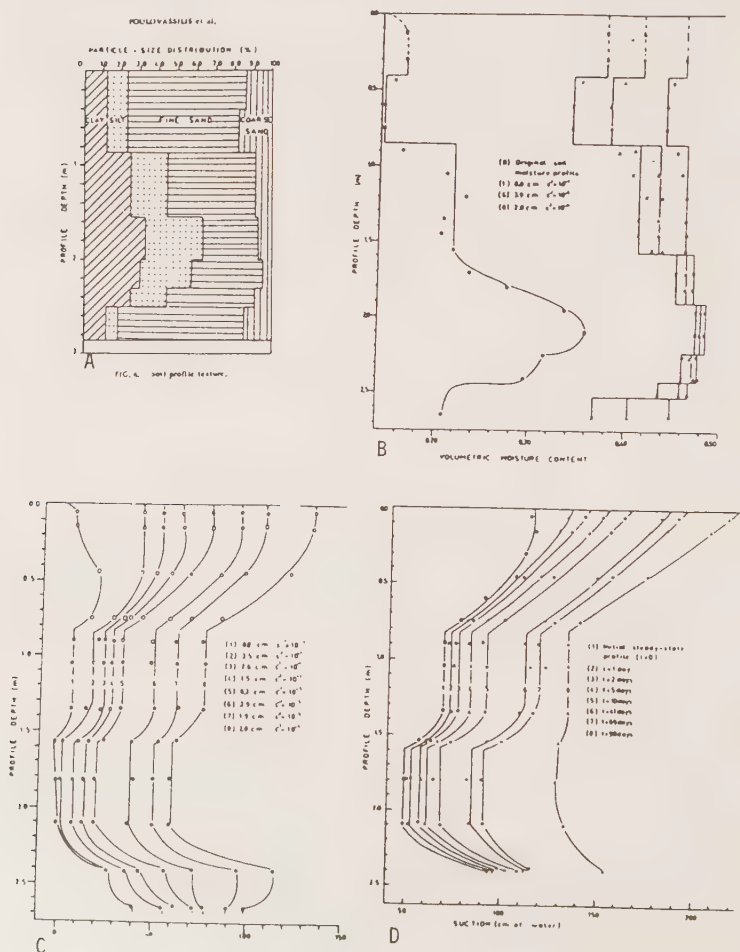
- a) Layered soils exhibit discontinuity of soil moisture profiles at the junctions of the different layers.
- b) The form of the discontinuity depends upon the sequence of the layering of the soil profile.
- c) In a fine-over-coarse layering the discontinuity is characterized by the shift of the moisture profile to the left while in a coarse-over-fine layering the shift is to the right, see figs. 16A and 16B

Ghuman and Pihar's experiments clearly show the effect of layering on tracer displacement. The tracer distribution in figs. 16A & 16B shows that the tracer is displaced deeper in the coarse-over-fine layering than in the fine-over-coarse layering.

The tail of the tracer peaks coincides with the wetting front in a coarse-over-fine layering and is widened and dispersed in a fine-over-coarse layering. When the wetting front reaches the boundary line in a fine-over-coarse layer, the high suction of the wetting front prevents it from crossing the boundary line and entering into the larger pores of the coarser material. Hence the high water content just above the junction where the tracer is sustained for some times. When the wetting front penetrates into the coarser layer, the bigger pores are first emptied in the upper layer to fill the smaller pores first in the coarser lower layer. This means that some of the tracer can still be remaining in the finer layer. This phenomena may explain the existence of double peaks encountered in the experiment.

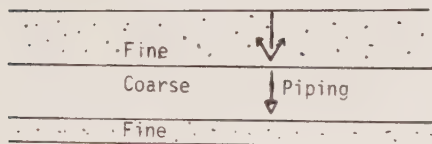
In a coarse-over-fine layer the redistribution of moisture is greater than in a fine-over-coarse layer. This is due to the higher retention of moisture in the finer layer and the resistance of the bigger pores in the coarser material to moisture entry from the finer layer.

Fig. 15 A layered soil profile (A) and the distribution of moisture (B) and suction (C and D), (From Poulovassilis et al, (1980))



In a fine-over-coarse layered formations, as shown in fig. 16C, because of the low conductivity in the coarse layer (under unsaturated conditions), the front of the percolating moisture can not immediately cross the bound-

Fig. 16 C. A fine-over-coarse layering



ary. This failure of the coarse layer to drain the upper layer causes a temporary saturation in the fine layer. This in turn initiates "piping" or a funnel-like flow at a particular point in the coarse material which can be slowed

when a fine layer is again encountered at a deeper level. This process may account for double tritium peaks. (E. Eriksson, personal discussion). These phenomena of double peak formation and hampered tracer displacement are also observed in the tritium experiment done by the author especially in Skediga area where coarser material underlies a relatively finer material near the surface.

Fig. 16 A and B. Stratification of soil and its effect on tracer displacement. (After Ghuman and Pihar, 1980)

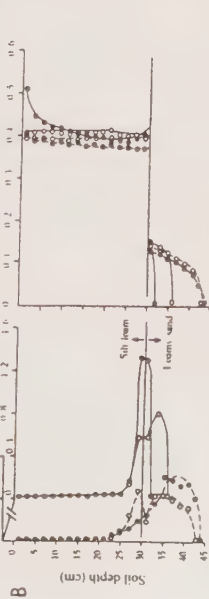
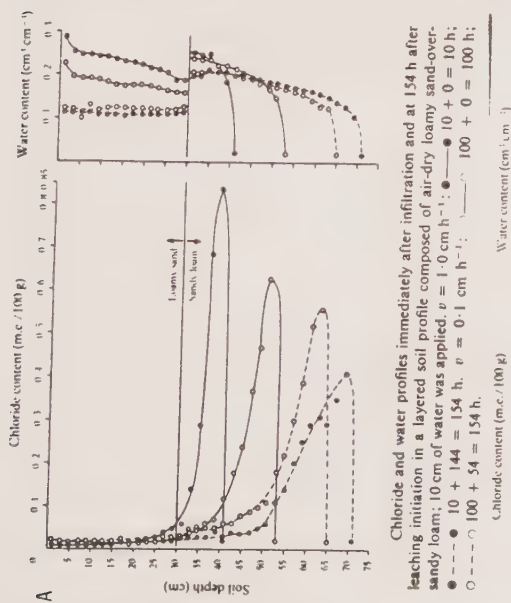


Fig. 4. As for Fig. 3, but for air-dry silt loam-over-loamy sand; 12 cm of water was applied. Note the shift in the scale of chloride for redistribution times to the left. Ponding: ●—● 11.5 + 0 = 11.5 h; ○—○ 11.5 + 156.5 = 168 h. $v = 0.1 \text{ cm h}^{-1}$; ○—○ 120 + 48 = 168 h.

4. VARIABLES USED AND THEORY OF FIELD MEASUREMENTS

The variables needed in this investigation are precipitation, which is the input part into the system in both the models; evapotranspiration, which is an important element in the water balance method and difficult to estimate accurately; soil moisture change with time and depth and tritium displacement in the vertical direction in a soil profile. The estimation or measurement of these variables is shortly discussed below.

4.1 PRECIPITATION (P)

Of all the different forms of atmospheric moisture, generally known as precipitation, rain (and / or snow) is the commonest form taking part in the hydrologic cycle. For Swedish conditions water from snow melt is the main source of recharge to ground water. In the measurement of precipitation, rain or the water equivalent of snow is expressed in units of length (usually mm). Precipitation is measured in the field either by non-recording or recording rain gauges (Rodda et al. 1976)

The main parts of the standard non-recording gauges are, the funnel shaped circular receiver with vertical sides and varying diameters, a cylindrical tube graduated on the side or with a measuring stick for finding the depth of precipitation. For example the U.S. Weather Bureau uses a circular receiver and a collecting cylinder of diameters of 8 inches and 2.53 inches respectively so that when one inch of rain falls in the receiver, the cylinder is filled to the 10 inches mark (Linsley, Kohler and Paulhus 1949).

The recording rain gauges also known as autographic or automatic rain recorders, usually work by means of a clock which drives a drum carrying a graph on which a pen records either the total weight or a series of blips for each time the container empties its content. Automatic recorders are either the weighing-type recording gauges or the tipping-bucket gauges.

Such recording gauges are more expensive and more liable to error (Wilson, 1969) but have the great advantage of indicating the intensity of rainfall and being applicable for remote and rarely visited sites. There are new tipping-bucket types provided with telemetric system for distant reading telephone interrogation. (Ibidem).

4.2 EVAPOTRANSPIRATION (ET).

The conversion of water from the liquid state to the vapour state takes place when solar energy falls on water surfaces, wet soil surfaces, snow and ice surfaces (evaporation) and on surfaces covered with plants and vegetation (transpiration). The two processes are known as evapotranspiration (ET). There are several ways of estimating evapotranspiration losses in nature, some of which are indirect methods while some are direct methods. Among the direct methods we have:

- (a) The pan or the tank method
- (b) The lysimeter method
- (c) Atmometer.

Some of the indirect methods are,

- (d) The water budget method
- (e) The energy budget method
- (f) The mass transfer method
- (g) Aerodynamic approach
- (h) The Thornthwaite formula
- (i) Crowe formula
- (j) The Turc formula
- (k) The Penman formula
- (l) The Penman-Monteith formula

A brief description of the different methods. (a) to (j), is given in appendix 11.4. The Penman method used in this investigation to estimate the potential evapotranspiration and the modified Penman-Monteith method are treated here.

(k) THE PENMAN FORMULA

The theory of Penmans method (1948) is based on the assumptions that;

- i) there must be a supply of energy to provide latent heat of vaporization and
- ii) there must exist some mechanism to remove the vapour produced.

The energy part is provided by solar radiation.

The former can be assessed by the energy balance approach and the latter by the aerodynamic approach. Combining the two approaches, Penman came to the expression,

$$ET = \left(\frac{\Delta R_n}{(\Delta + \gamma)} + \frac{\gamma E_a}{\Delta + \gamma} \right) \frac{1}{\lambda} \quad (38)$$

where,

ET = potential evapotranspiration

Δ = slope of the saturation vapour pressure curve versus temperature

R_n = the net solar radiation received by the surface

γ = psychometric constant (0.66 mb/ $^{\circ}$ C)

E_a = the aerodynamic term

λ = latent heat of vaporization of water

The energy part (R_n) and the aerodynamic part (E_a) can be expanded in the following manner.

$$R_n = (1 - a)R_s + R_l - \epsilon \sigma T^4 \quad (39)$$

$$E_a = 0.26 (0.5 + 0.54u) (e_s - e) \quad (40)$$

Therefore,

$$ET = \frac{\Delta \{ (1 - a)R_s + R_l - \epsilon \sigma T^4 \} + \{ 0.26(0.5 + 0.54u) (e_s - e) \} \gamma}{(\gamma + \Delta)} \frac{1}{\lambda} \quad (41)$$

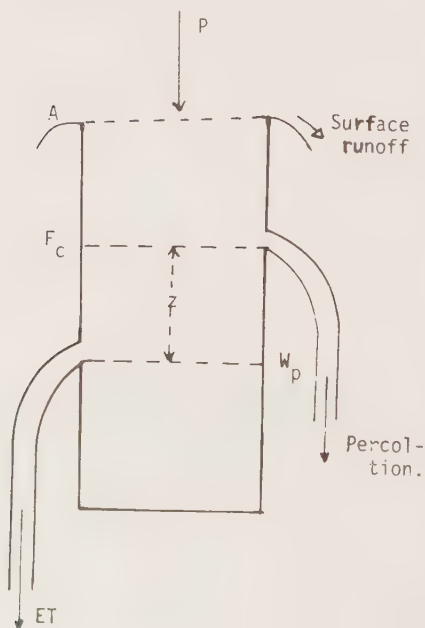
In equation (41),

- a = reflection of the earth's surface or albedo
- R_s = Incoming solar radiation
- R_l = longwave radiation
- ϵ = emissivity
- σ = Stefan-Boltzmann's constant
- T = absolute temperature in $^{\circ}\text{K}$
- e = actual vapour pressure of the air
- e_s = saturation vapour pressure of the actual temperature
- Δ = de_s/dt in mb per $^{\circ}\text{C}$
- γ = adiabatic lapse rate
- u = wind speed in m/s

In the study of drainage areas under conditions of unlimited supply of water, one speaks of the potential evapotranspiration (PET) and under conditions of limited water supply during a certain period of the year, (for example when soil moisture condition is the limiting factor) one speaks of the actual evapotranspiration (AET). The soil moisture content in the

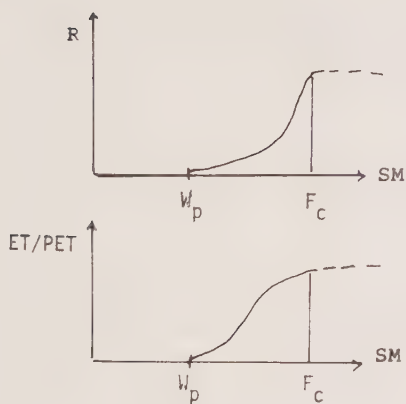
unsaturated zone is conceptually described by the simple diagram, fig.17A (Eriksson, E. 1972, comp., unpublished). The soil profile is considered to be a beaker with the three levels which describe the three hydrological concepts essential in soil moisture studies. When the infiltration rate is exceeded by the intensity of rainfall, some of the precipitation becomes runoff (level A). As infiltration proceeds, the soil moisture content increases, as a result of which the gravitational force exceeds the capillary potential. At this point water percolates into the lower zone. This point at which gravitational flow starts and ceases

Fig.17A Soil moisture storage



or the water content at which the pF value is equal to 2.2 is called "the field capacity" (F_C). When the soil moisture content decreases to the point where the pF value is equal to 4.2 and impossible for plants to extract for growth, this point is a wilting point (W_p). Evaporation

Fig.17B A graph of percolation and evaporation versus soil moisture.



loss (ET) is possible only as long as the water content is above the wilting point since at this point the moisture in the soil is no longer available for plant use. The soil moisture content between the field capacity and the wilting point, "z", is of hydrological importance. Under natural conditions the relation of percolation and evaporation rates to soil moisture content is non-linear as shown in fig.17B. Therefore since almost all the methods for the estimation of evaporation are based on unlimited supply of moisture and hence give the potential evaporation (PET), these values must be converted

to the actual evapotranspiration values by using a certain function which depends on the soil moisture conditions. One such relationship is shown in the equation below (Dunne and Leopold, 1978).

$$ET = PET \left\{ \frac{(\theta - W_p)}{(F_C - W_p)} \right\}^\beta \quad (42)$$

$(\theta - W_p)$ = the available soil moisture

$(F_C - W_p)$ = the available water capacity of the soil

θ = the actual soil moisture content

β = a constant

According to Eriksson and Johansson (1978) the best estimate of the actual evapotranspiration is obtained when the constant β is equal to 0.5.

(1) THE PENMAN-MONTEITH FORMULA

For vegetated or forest areas, the Penman formula can be modified to take account of the resistance to flux between the stomata and the leaves (surface resistance, r_s) and between the surface of the leaves and the open air (aerodynamic resistance, r_a). Monteith introduced these factors (Rodda et al, 1976) in the penman formula. The Penman-Monteith formula is written as,

$$ET = \frac{1}{\lambda} \cdot \frac{\Delta R_n + \rho C_p \cdot \{(e_s - e)/r_a\}}{\Delta + \gamma(1 + r_s/r_a)} \quad (43)$$

where,

$$r_s = \frac{\rho C_p}{\gamma} \cdot \frac{e_s(T_0) - e_0}{\lambda E} \quad (44)$$

$$r_a = \frac{\rho C_p}{\gamma} \cdot \frac{e_0 - e_z}{\lambda E} \quad (45)$$

r_s is Monteith's canopy resistance to the transport of water from some region within or below the evaporating surface to the surface itself and is expected to be a function of the stomatal resistance of the individual leaves.

r_a is the aerodynamic resistance to the transport of water vapour from the surface to the reference level z .

ρ = density of air; e = vapour pressure in mb

e_s = saturation vapour pressure

C_p = specific heat of air at constant pressure

$e_s(T_0)$, e_0 = saturation vapour pressure at temperature T_0 and vapour pressure of the canopy air.

e_z = vapour pressure at height z

The other symbols are as given in equation (41).

The main advantage of the equations (43) - (45) is that they can be applied by using data on meteorological variables but have one big disadvantage and that is the difficulty of adequately measuring the factors r_s and r_a . When the surface resistance r_s is equal to zero, then equation (43) reduces to the Penman equation for potential evapotranspiration and when the aerodynamic resistance r_a is equal to zero, it gives the evaporation loss from a wet vegetated surface. Depending upon the vegetation and other factors, the ratio r_s/r_a usually takes the value of about 30 and typical values of r_s and r_a are 100 and 3 respectively. (Björn Bringfelt, 1979, unpublished)

4.3 SOIL MOISTURE DETERMINATION

The simplest and traditional way of determining soil moisture content is to take soil cores dry them and compare the moisture loss in weight with the weight of the soil. For soil moisture estimation on a routine basis many types of in situ soil moisture measurements have been tried by many workers. Some of these methods are discussed below.

(a) THE GRAVIMETRIC METHOD

In this method soil cores are collected and weighed in the laboratory before and after heating them at 105°C for 24 hours. This gives the fraction of water to the soil material by weight. To change the weight fraction into units of length (Eg. mm), the moisture content must be expressed by volume fraction. This necessitates the determination of bulk densities. Another way of getting soil moisture by volume fraction is to take samples with cylinders of known volume (usually 5 or 10 cm high). The main disadvantage of this method is that the coring is destructive to the sampling site and consequently the temporal and spatial changes of soil moisture content are difficult to follow. In this investigation, the method is used for calibration purposes only.

(b) THE ELECTRICAL METHOD

The electrical method is based on the principle of the change of the electrical properties of an electrode inserted into the soil with the soil water content. The first attempt to use the electric conductivity method by Whitney et al. (1898) was not of any success because the soil electric conductivity is more dependent on its salt solutions than the moisture content. Shaw and Baver (1939b) used the dependence of heat conductivity in soils on its moisture content to estimate soil moisture content. This method can measure soil water over the entire moisture range. Croney et al. (1951) used the electrical resistance units for the first time in Great Britain. In this method two electrodes are inserted into the soil profile. The resistance which is dependent on the moisture content of the soil is measured by a Wheatstone Bridge. The main sources of error are temperature and aging. Calibration difficulty is another drawback of the method.

(c) TENSIOMETERS

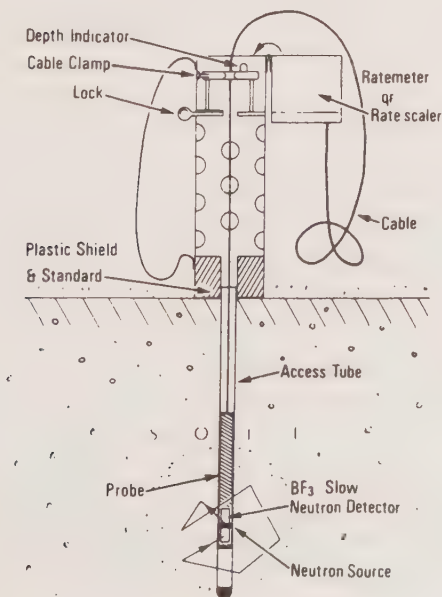
In this method ceramic porous cup attached to a tube is placed in the soil. The tube is provided with a manometer. The porous cup and a portion of the manometer tube are filled with distilled water. As the soil gets drier water leaves the porous cup and enters the soil and when the soil becomes wetter the potential in the soil falls and water enters into the porous cup. When the waters in the porous cup and in the soil are in equilibrium, the manometer reading gives the potential of the soil. These tension values measured by tensiometers are converted into moisture content. The suction method measures moisture at the wet end of the energy-moisture curve (ie. 0 - 0.85 atms.) because at higher suctions air enters into the system. However, since this range includes over 50% of the available water from field capacity to the permanent wilting point for fine-textured soils and about 90% for sandy soils (Richards and Wadleigh, 1952) the tensiometer method is a useful tool in soil moisture studies. For tensions higher than 1 atms. the pressure membrane apparatus is used (Richards, 1949b).

One other drawback of the tensiometer method is the hysteresis effect which gives different manometer readings depending on whether the soil is drying or being wetted.

(d) THE NEUTRON MOISTURE PROBE

For routine soil moisture change determinations at a given site, the neutron moisture probe is a very convenient instrument. The main components of the instrument are the source, the detector and the counter (see fig. 18). The principle of the neutron probe method is based on the fact that fast neutrons produced by the (α, n) reaction in a source (Re - Be or Am - Be source) are slowed down by the elastic collision with the hydrogen atoms mainly present in water molecules.

Fig. 18 A sketch of a neutron probe (taken from Rodda et al., 1976)



The source is sunk to the desired depth by means of an access tube and cable. Am - Be source is a widely used source and the reaction reads,



In this type of source the berillium is bombarded by the α -particles from americium. Some of the fast neutrons undergo elastic collisions with the hydrogen nuckli of the water molecules and acquire a lower velocity. The density of the slow neutrones thus formed is a direct measure of the water molecules. The detector in the probe is mostly sensitive to the slow or thermal neutrons. The proportionality of the thermal neutrons to the water content is not, however, constant. Factors such as soil bulk density, mineral contents etc. influence the results. Hence, each site needs a separate calibration curve for the determination of soil moisture content corresponding to the neutron counts. The count rates are transformed into volumetric moisture content by a relation of the following form,

$$\theta = A \cdot B + M$$

where, θ = volumetric water content

A = slope of the calibration curve

B = ratio of count rate in soil (B_s) to count rate of standard or reference (B_r) = (B_s/B_r).

M = calibration constant (the negative intercept on the θ axis).

A and M depend upon the property of the soil. Calibrations can either be done in the laboratory or in the field. To perform a field calibration (which is more accurate than the laboratory calibration or than the curve provided by the manufacturer), soil cores of known volume are collected from around the access tube (temporary) at each depth. The moisture content by volume fraction is determined by the gravimetric method and these values are plotted against the count rates. This is repeated for different soil moisture conditions.

In the laboratory method, a big drum is filled with a soil material and the neutron probe is used to compare the count rates at different soil moisture contents. This method is tedious, difficult and unreliable. The neutron probe method gives a non-destructive and quick way of determining of soil moisture changes.

(e) THE TDR (TIME DOMAIN REFLECTOMETRY) METHOD

The TDR method of soil moisture measurement exploits the possibility of using the dependence of the complex dielectric constant (K) of the soil on its volumetric moisture content. Extensive laboratory studies have proved that the dielectric constant of soil varies very little with its density, texture, salt content or temperature (Thomas, 1966; Lundien, 1971; Cihlar and Ulaby, 1974; Okrasinski et al., 1978; Topp et al., 1980).

To measure the dielectric constant of a soil in the field, the following procedures are performed. First a pit is dug to the required depth and pairs of stainless steel rods are inserted into the soil. The two rods with the soil material in between make up a system of capacitance. (see fig. 19). The static suppressor connected to the TDR unit is meant for protection of the instrument against static electricity that may build up.

When the TDR unit is switched on, a pulse of electromagnetic wave is sent out through the transmission lines. The propagation of an electromagnetic wave is described by the equation (Topp et al., 1980a),

$$v = c/\sqrt{K} \text{ or } K = (c/v)^2 \quad (48)$$

where,

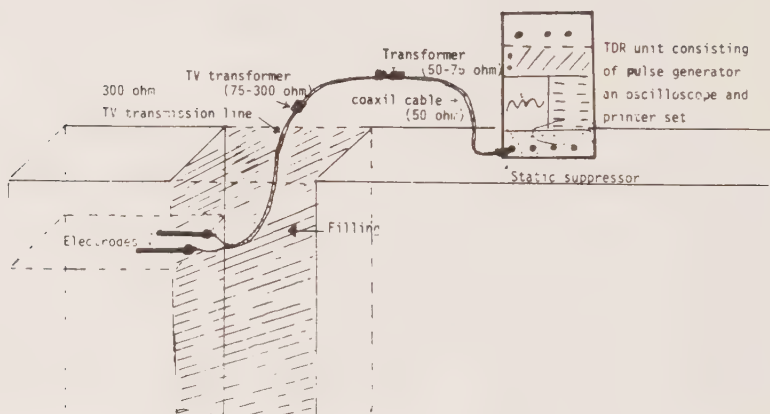
K = the dielectric constant

c = speed of light

v = velocity of propagation

In equation (48) the two unknowns are K and v . The time domain reflectometer measures v to calculate the K values.

Fig. 19 A schematic diagram showing a set up of TDR method in the field



Using a transmission line of known length, L_1 , then,

$$K = (ct/L_1)^2 \quad (49)$$

As the electromagnetic pulse passes through the electrodes, there happens a disturbance that is characterised by a sudden jump at the beginning of the probes and a steady rise at the end of the probes. This trace is displayed on the oscilloscope, the size and shape of which can be controlled by means of coordinate tangents on the instrument. A typical trace is shown in fig. 20. The exact position of the two ends of the trace is found by drawing a perpendicular to the peak at the beginning and tangents at the end of the trace. If the length of the trace on the oscilloscope is L_0 , then, the velocity of propagation v can be expressed as,

$$v = L_0/t \text{ and } t = L_0/v \quad (50)$$

Substitution of equation (50) in equation (49) and setting $v = c$ gives,

$$K = (L_0/L_1) \quad (51)$$

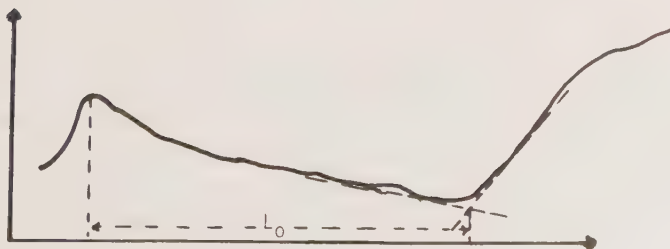
The dielectric constant thus obtained is converted to volumetric soil moisture content by using a calibration equation of the following form, Topp et al. (1980a).

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} K - 5.5 \times 10^{-4} K^2 + 4.3 \times 10^{-6} K^3 \quad (52)$$

The soil moisture contents measured by the TDR method and the gravimetric method were compared by Topp et al. (1980b) and the discrepancy was $\pm 0.03 \text{ cm}^3 \text{ cm}^{-3}$. In another lab experiment they found that the average volumetric water content deviated by less than $0.009 \text{ cm}^3 \text{ cm}^{-3}$. Jean Stein and Douglas L. Kane (unpublished) have successfully used the TDR method to measure the unfrozen soil moisture content in freezing soils and to study the infiltration of snow melt. To conclude the TDR method seems to be a promising method in soil moisture studies especially in freezing soils.

The TDR method is being tested in Uppsala at two sites, one in a clayey soil near Ultuna and another in a sandy soil near Ulleråkers, the results of which shall be presented in a forthcoming report.

Fig. 20. Typical trace: Finding beginning and end of line.



4.4 TRITIUM PROFILES

In order to obtain tritium profiles, tritiated water is injected into the soil profile, preferably under the root zone. Assuming that soil moisture moves in layers, the advance of the injected tritium peak is followed. In a given homogenous soil profile, the displacement of a tracer that does not react with the soil materials (such as tritium), is described by the following equation.[66,78,79]

$$\frac{\partial C}{\partial t} = D \frac{\partial^2 C}{\partial x^2} \quad (52a)$$

where,

- C = concentration of tracer
- t = time
- x = the distance along the column
- D = coefficient of dispersion analogous to an apparent mol. diff. coef.

Assuming constant average pore water velocity, v , the solution of the above equation gives:

$$\frac{C}{C_0} = \frac{1}{2} \left\{ \operatorname{erfc} \left(\frac{z - vt}{\sqrt{4Dt}} \right) - \operatorname{erf} \left(\frac{z + z_0 - vt}{\sqrt{4Dt}} \right) \right\} \quad (52b)$$

C, v , t , and D are as given above

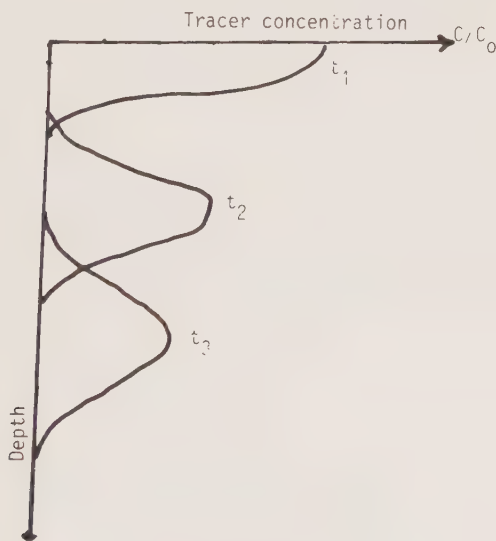
C_0 = Initial concentration

z_0 = the depth the tritiated water occupies within the profile.

As we see from equation (52b), the distribution of tracer, C/C_0 is a function of time and depth. For some fixed times t_1 , t_2 , and t_3 , the tracer distribution C/C_0 as a function of depth and as described by equation (52b) gives a Gaussiann distribution as shown in the figure 20a. In the field where soil profiles are usually heterogenous, the tritium distribution is mainly non-Gaussian, but the peak concentration

can frequently be identified. In cases where this is impossible the center of gravity (CG) is used instead of a peak. The usual procedure during field work is, when enough time has elapsed after the tracer injection, usually a month or two, soil moisture samples (by coring, tensiometer or suction method) are collected for different depth intervals. The tritium concentration of the water samples is determined in the liquid scintillation spectrometer and the results in cpm or TU are plotted against depth. The graph gives the tritium profile required.

Fig. 20a A theoretical tritium profile.



5. MEASUREMENT TECHNIQUES USED

5.1 PRECIPITATION

For Skediga site, precipitation in the form of rain is measured using the "OTA" type recording rain gauge. The rain gauge was placed in an opening near the sampling stations at 150 cm height. The measured values at Skediga show a very good agreement with those measured at the military meteorological station F16 which is situated two to three Km from Skediga. Precipitation measurements at Skediga are done for the summer months only and the missing values are completed using the data from F16. The precipitation values for Jädraås and Tärnsjö are obtained from data collected by the "Swedish Coniferous Forest Project" and the "Swedish Geological Survey" (SGU) respectively.

5.2 EVAPOTRANSPIRATION

The potential evapotranspiration for Skediga site is estimated by using the Penman formula. The meteorological data for the Penman equation were obtained from the station at F16. The potential evapotranspiration (PET) is used to estimate the actual evapotranspiration (ET) by using the following equation.

$$ET = PET \times \left(\frac{\theta - \theta_{\min}}{\theta_{\max} - \theta_{\min}} \right)^{\beta} \quad (53)$$

where,

θ = the actual soil moisture content measured

$\theta_{\max}, \theta_{\min}$ = the maximum and minimum soil moisture contents of the period.

β = constant

A value of 0.5 was given to the constant.

The necessary evapotranspiration values for the estimation of recharge in Jädraås area were obtained from Per Erik Jansson who also provided the soil moisture data needed.

For Tärnsjö area, where no meteorological data are available, data from Ljusbäck, some kilometers away is used.

5.3 SOIL MOISTURE MEASUREMENT

The soil moisture content in all the three sites is monitored by the neutron scattering technique. In Tärnsjö area the Danish type equipment with a lithium-glass scintillator and photomultiplier as a detector and 50 mC Am-Be source is used. The measurements for this site are done by Thomas Aneblom from the Swedish Geological survey (SGU), while Per Erik Jansson from the Swedish Coniferous Forest Project did the measurements for the site in Jädraås. For Skediga and Jädraås areas the Wallingford Neutron Moisture Probe is used. This instrument is very light (10kgs.) and easy to carry in the field from site to site. The source is made up of 50 mC Americium-Meryllium while the detector is a standard 20th century 12 EB 70/25/G borontrifluoride ($^{10}\text{BF}_3$). The instrument has a count rate of 1000 cps in water and 5 cps in air. A pack of 12 volt battery supplies the power required to activate the scaler. It has a carrying strap on the transport shield which makes it easy to carry the complete instrument on the shoulders from station to station. In Skediga, access tubes of 1 1/4 inch stainless steel tubes were installed with a projection of about 50 cm above the ground surface. Two access tubes, a meter away and on each side of the injection plates were positioned. A lid with a screw lock was made for each access tube. Soil samples were collected using 10 cm cylinders for gravimetric soil moisture determinations. Using the field method discussed in section 4.3 (d), a calibration curve was made. The following calibration equation is used to convert the neutron counts to volumetric moisture content.

$$\theta = -3.4 + 50 \cdot B \quad (54)$$

where the units are as given in equation (47).

5.4 THE SOIL PROFILE STUDIES --

The soil profiles in Jädraås were studied by Per Erik Jansson (1977) and in Skediga by the author. Using 10 cm cylinders, soil samples were taken at 10 cm intervals along the depth and the grain size distribution analysis was performed.

5.5 TRITIUM MEASUREMENTS IN THE FIELD

5.5.1 SELECTION OF SAMPLING POINTS

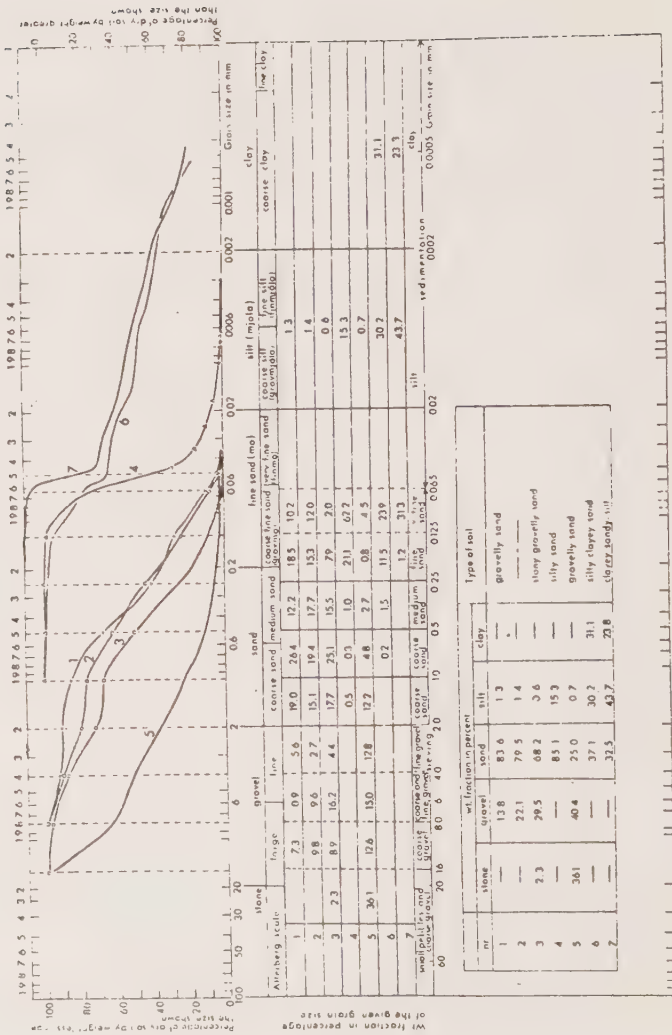
When this project started in the summer of 1976, a thorough reconnaissance study of the Skediga area was done. This area was in the first place selected mainly for three reasons. namely;

- (a) Its preferable location about 10 km from populated area
- (b) It is part of the esker formation in the area (Uppsala esker) which as a glaciofluvial deposit is of prime interest for the study of ground-water formations
- (c) The presence of the meteorological station (F16) in the vicinity of the site and the possibility of obtaining some geological and hydrological data from the municipality of Uppsala town (Sidenvall, 1975).

In order to have the sampling points at representative places, areal surface-soil (0 - 70 cm) analysis was done. Point samples at randomly selected spots were taken and grain size analysis was performed. The results of the grain size distribution of the surface soil are shown in fig. 21. Based on this analysis the area was schematically divided into the different soil types as shown in fig. 22 on which the point-sampling spots are also indicated. The position of the plots for the experiment was then selected in such a way that the different soil types are included. The grain size distribution analysis clearly shows that the particle size increases towards the ridge of the esker in a NW to SE direction.

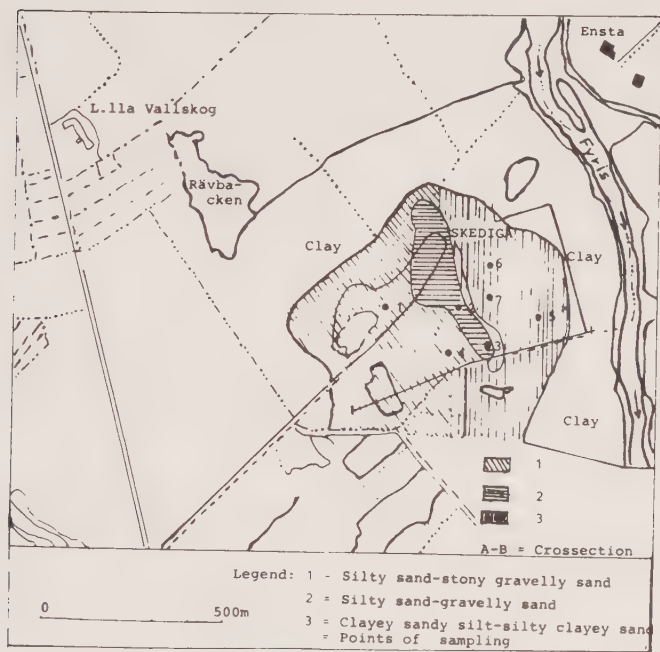
Tritium injection plots of 1 m^2 were selected at sites representing the different soil types. Glaciofluvial deposits making esker formations undergo a process of sorting by which the heavier materials such as gravels and stones form long winding ridges with the lateral parts of finer material, whose grain size decreases with the distance from the center of the ridge. In this area the sandy and the gravelly areas are represented. Accordingly plots A and B are in the sandy area while plots C and D are in the coarser material, D being on the south east edge of the ridge. Plot E is in the clayey area and is abandoned because sampling by the porous cup method failed to work.

Fig. 21 Grain size distribution for the top 70 cm of the soil profile, Skediga



Also, the vertical rate of flow in such materials is assumed to be too low to be monitored at the same rate as in the coarser materials.

Fig. 22 A sketch showing the surface soil mapping.



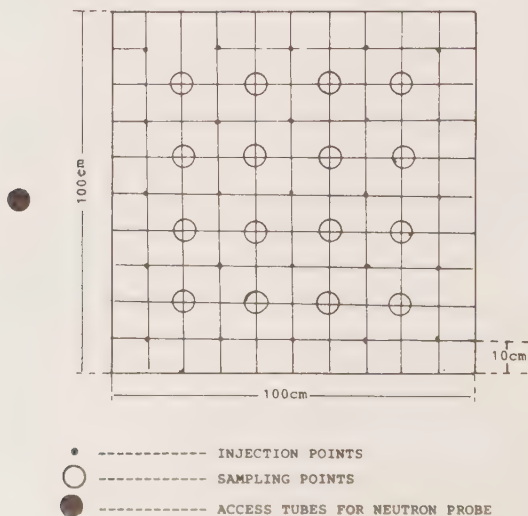
The same type of procedure in reconnaissance study was not performed for the other two sites at Jädraås and Tärnsjö. The selection of the injection points in Jädraås was primarily based on the vicinity of the plots to the existing installations of tensiometers and access tubes for the neutron soil moisture probe. Out of the five installations only two were used due to the relatively low soil moisture content and difficulty in obtaining enough samples for analysis from all plots.

The last site that came into the picture was at Tärnsjö. The geological and hydrological studies of the area are done by the Swedish Geological Survey (Norberg and Persson, 1974). No soil profile studies are done for this site. The selection of the injection plots was based mainly on the geological description available and also on the vicinity to observation wells. The location of the experimental plots is shown in fig. 4. A recording rain gauge and a piezometer (which is also used as access tube to the soil moisture probe) are installed near plot C. Soil moisture measurement lacks for the other plots. According to the geological information, plots A, C and E are in sandy and very coarse sandy formations while plots B and D are in gravelly formations.

5.5.2 METHOD OF INJECTION

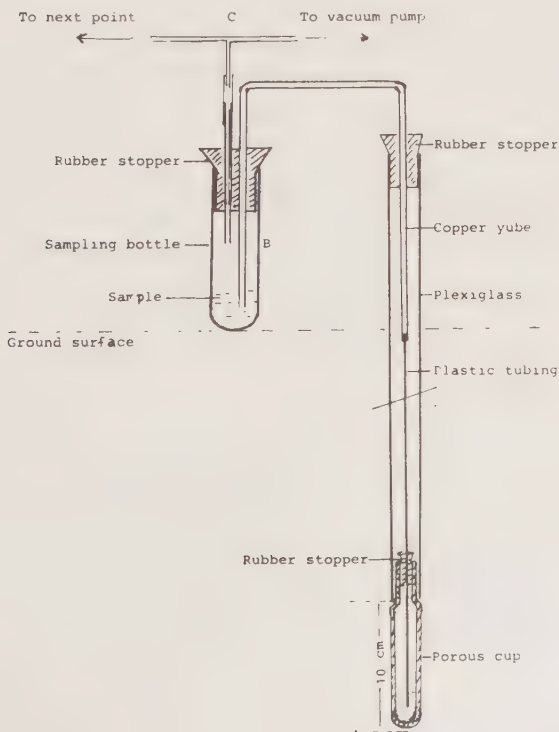
The injection plots are 1 m^2 in area. A wooden frame of the same size was made. The area inside the frame was divided into smaller areas of 10 cm^2 by means of strings fixed to the sides of the frame like a net-work. (see fig. 25 below.)

Fig. 25 A plan for injection and sampling tritium



16 points 20 cm apart (shown with open circles) are used for installing the porous cups (sampling points). The four corners round each sampling point make up injection points (shown with black dots). Each sampling point was made by first making a hole with an iron tube of an outer diameter equal to that of the porous cups ($3/4$ inch). The iron tube is slotted on the sides to make it easy to empty it of the soil material. The porous cup which is glued to one end of the 22 mm plexi-glass tube (fig. 26) is inserted into the hole made to the required depth. The sides of the sampling tube are then packed with the same soil material to prevent direct side-infiltration.

Fig. 26 Sampling arrangement



According to the specifications by the manufacturer, the flow rate of water through the cups is approximately 5 ml/min at a pressure of 1 bar differential across the cup wall.

The first sampling depth is 55 cm from the surface and 15 cm from the injection depth (40 cm from the surface). The successive sampling depths are at 15 cm intervals up to the depth of 160 cm (i.e. 8 levels). Each level has a duplicate as a precaution in case sampling fails in any one of the sampling tubes.

For injection, holes are made using a piece of metal rod of about 10 mm diameter up to 40 cm depth. A copper tube of a smaller outer diameter (5 mm) is used for injection. An exact amount of tritiated water (10 ml in this case) is obtained by using a sampler and syringe attached to a bottle. The permitted concentration of tritium, as prescribed by National Institute For The Protection Against Radiation was 10 μ Ci per plot. After injecting the 10 ml of tritiated water, any remnant in the injection copper-tube was removed by means of a hand pump.

The high density of injection points was meant to ensure a uniform sheet movement of the tracer front. It should be pointed out that the symmetry of both the injection and sampling points can be disturbed because of the installation and injection problems caused by very coarse materials. After installation of the sampling tubes, a period of two to three weeks was allowed to elapse before a background sampling is done to find out the natural or the background tritium content in the profile. Only then is injection performed.

5.5.3 INJECTED-TRITIUM SAMPLING METHOD

Each sampling tube is connected to a sampling tube (B) by means of a three-way connection tube (C). (see fig . 26). All the sampling tubes are connected in parallel to the pumping end.

In order to have water in the porous cups for sampling, moisture must be extracted from the soil surrounding it. This is done by creating a pressure gradient between the outer and the inner walls of the porous cups. A small vacuum pump that is run by (2 x 12V) car batteries is used to achieve this pressure gradient. The whole system is left under the required pressure gradient for some hours (sometimes overnight) to allow enough water to collect in the porous cells. Then a pressure push is applied to suck out the moisture sample into the sampling tubes. Any moisture that existed in the porous cups is sucked out before the sampling process is started. The samples in the sampling bottles are then transferred into 25 ml brown bottles for further transport for laboratory analysis.

5.5.4 TRITIUM ANALYSIS - TRITIUM LABORATORY

5.5.4.1 INJECTED TRITIUM

The tritium activity in the samples is determined by using the Packard Tri-Carb Liquid Scintillation Spectrometer, Model 3002. This instrument has an automatic sample changer which can accommodate 200 samples at a time. The main parts of a scintillation spectrometer are, (a) high voltage supply, (b) photomultiplier tubes (PMT), (c) a refrigerator, (d) pulse discriminator, (e) a scaler and (f) coincidence circuit (fig. 27).

(a) A high voltage supply is indispensable to obtain a stable gain from the photomultiplier tubes.

(b) The photomultiplier tubes collect light flashes produced when the ions from the radioactive substance cause excitation and ionization of the solvent molecules which in turn transfer their energy to the fluorescent substances that fluoresce or scintillate.

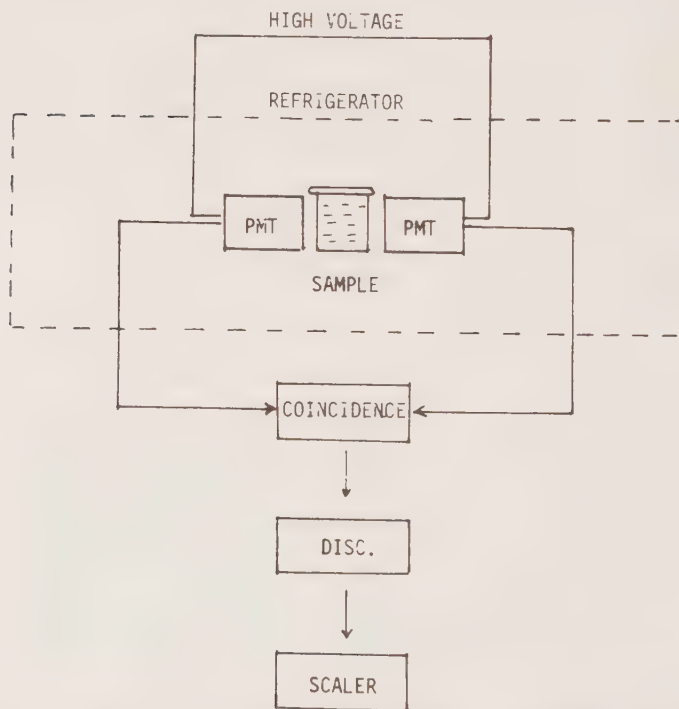
(c) The refrigerator unit in which the detector unit is placed is meant to prevent thermoionic emission which can increase the background counts.

(d) The pulse discriminator eliminates all other pulses than those which correspond to the nuclear event in question.

(e) The scaler transforms the pulse rates to counts per minute (cpm) which is printed on paper by the special printer in the instrument.

(f) The coincidence circuit is used to produce one output pulse if it receives an input pulse from each photomultiplier in coincidence.

Fig. 27 block diagram of a liquid scintillator



Before counting is done an equal volume of sample and liquid-scintillation mixture (instagel in this case) are transferred into special counting vials made of linear polyethylene and obtained from Nuclear Chicago Corporation (PEN) or from Packard Instruments (PEP). A commonly used liquid-scintillation mixture contains 4 g of PPO, 0,05 g of POPOP and 100 g of naphthalene/l of dioxane. PPO (dipheniloxazole) is the primary solute and POPOP (bis-phenyloxazoly1-benzene) is the secondary solute.

The ionizing particles from the radioactive material excite and ionize the solvent (dioxane) molecules which transfer their excitation energy to the PPO molecules which in turn fluoresce or give rise to light photons (scintillate). The secondary solute (POPOP) has two purposes, one is the remission of the absorbed light in a wavelength that is more suitable for the phototubes and the second is to reduce the effect of quenchers.

The water sample mixed with the liquid-scintillation mixture is put in the scintillation spectrometer for counting. The result is printed in counts per minute (cpm).

To be able to convert the counts per minute to tritium units (TU), a standard sample with known TU contents and a background sample (dead water) are run with the samples to be analysed. If N_t , N_b and N_s are the count rates of the standard, the background and the water sample, and if N_{tu} is the known tritium content of the standard, then the tritium content of the sample is:

$$TU = \frac{N_s - N_b}{N_t - N_b} \times N_{tu} \quad (55)$$

$$1 \text{ TU} = 1 \text{ } ^3\text{H} / 10^{18} \text{ } ^1\text{H} \text{ (one tritium atom per } 10^{18} \text{ atoms of hydrogen)}$$

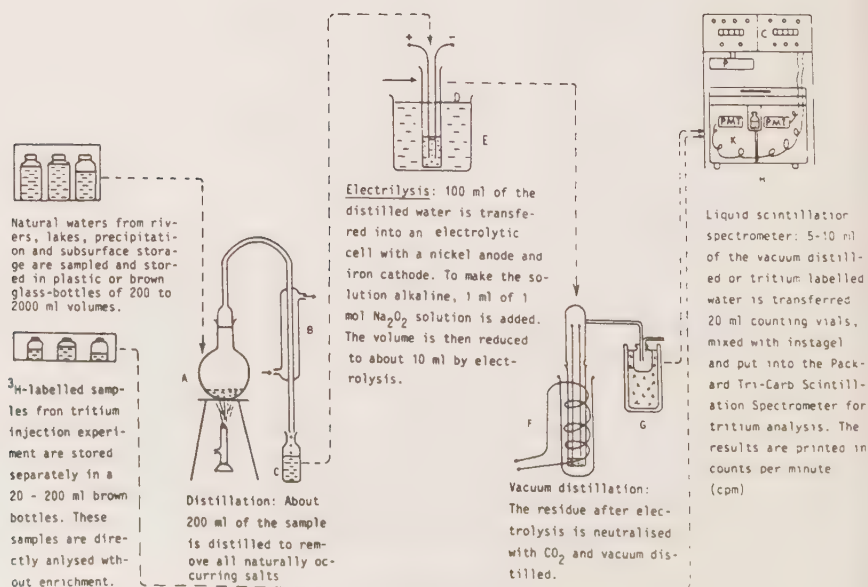
5.5.4.2 NATURAL TRITIUM

The tritium concentration in precipitation in the northern hemisphere reached its maximum of a few thousand TU in 1963 and since then it has continuously been falling. The present concentration in precipitation is far below 100 TU. (Florkowski, 1981). In most natural waters, specially in subterranean waters the tritium concentration is nearing the natural pre-bomb concentration of about 5 - 10 TU. Because of this very low concentration and very low energy (18.27 keV) natural waters must be treated in a special way which calls for an enrichment process.

The tritium lab is also equipped with an enrichment plant which is the only one in Sweden at present. The electrolytic enrichment being applied is similar to that used by Brown and Grummitt (1956) and Østlund and Werner (1962). The procedure of the enrichment process is schematically shown in fig. 28.

The principle of the electrolytic enrichment is based on the fact that electric current is allowed to circulate in the sample whereby $^1\text{H}_2^{16}\text{O}$ molecules are dissociated in preference to the heavy $^3\text{H}_2^{16}\text{O}$ molecules. This increases the ^3H -concentration in the reduced volume of the sample thus making detection easier. Tritium labelled waters from tritium injection experiments need not be enriched because of the high activity.

Fig. 28 A schematic representation of the activities in the tritium lab.



Natural waters contain differing amounts of salts. Since conductivity depends on salt content the electrolysis may not go at the same rate in all the samples. Therefore all naturally occurring salts are removed from all samples by distillation. About 200 ml of the sample is transferred into the round-bottomed flask (A) and boiled by a gas burner. The water vapour is cooled by running water (B) and trapped in the brown bottle (C).

Before electrolysis is started, some electrolytes are introduced to make the conduction of current easy. Electrolyte concentration of 1-2% is considered as the lower limit and 20% as the higher limit (Theororsson, 1973). In our case we use sodium peroxide as a source of electrolytes.



The electrolysis cell with the sample is immersed in a cold bath to avoid evaporation loss. The initial current of 3.0A is reduced to 1.5A after 24 hours and still lower, 0.3A, after a week and half. A volume reduction from 100 ml to 10 ml takes about 3 - 4 weeks.

When the electrolysis is ready, the samples are recovered by vacuum distillation. This is done by heating the cells to about 100 °C and trapping the vapour in a cold bath of a mixture of CO₂-ice and 95% spirits. Before vacuum distillation the enriched sample is neutralized by passing carbondioxide through it. This is done to prevent large amounts of hydrogen from being bound to sodium hydroxide.



The Østlund type of enrichment set up discussed above takes a relatively long time to electrolyze a batch of samples with 10% volume reduction. A more rapid type of electrolytic cell as used in Vienna and by Theodorsson is being tested in this laboratory also. The cell is made up of an inner nickel anode and an outer stainless steel cathode in a cylindr-

ical form. The cells, 10 in number are immersed in a cold bath of about 0 °C and a current of 10 A is passed through them. The electrolysis takes only two days per batch. This increases the capacity of the enrichment plant which is the only one of its type in Sweden. By this method a volume reduction from 100 ml to 1 ml can easily be achieved. At present the method is not being used for routine measurements due to lack of financial support.

The important terms in connection with enrichment processes are "recovery" "enrichment factor" and "enrichment parameter".

If MT_0 , MT , MD_0 and MD are the original and final molar concentrations of tritium and deuterium then according to Brown and Grummitt (1956),

$$MT/MT_0 = (MD/MD_0)^{\mu/\omega} \quad (58)$$

where, μ = separation factor for deuterium

ω = separation factor for tritium

μ/ω = enrichment factor

According to Kaufman and Libby $\omega/\mu = 2.10 \pm 0.10$

If V_0 and V are the initial and final volumes of the sample and MT_0 and MT are as given above, then according to Østlund and Dorsey, the recovery, RT , can be expressed as follows:

$$RT = VMT/V_0MT_0 \quad (59)$$

The "enrichment parameter", EP , is introduced by Tayler (1978) and is expressed in the following manner,

$$EP = \{(2.975(W_0 - W))/J\} \cdot \{\ln(EF)/\ln(W_0/W)\} \quad (60)$$

where,

J = number of ampere-hours during the electrolysis

2.975 = number of ampere-hours for the electrolysis of 1 g of water

W_0 , W = initial and final mass of water (g)

EF = enrichment factor for tritium

EP is related to the recovery of tritium by the expression.

$$EP = (W_0/W)^{-1/\omega} \quad \text{where} \quad -1/\omega = EP - 1 \quad (61)$$

6. THE HYDROGEOLOGY OF THE INVESTIGATION AREAS

The hydrogeological description of the experiment areas is based on the existing data obtained from the different works available. [37,47,59,60]

6.1 JÄDRAAS

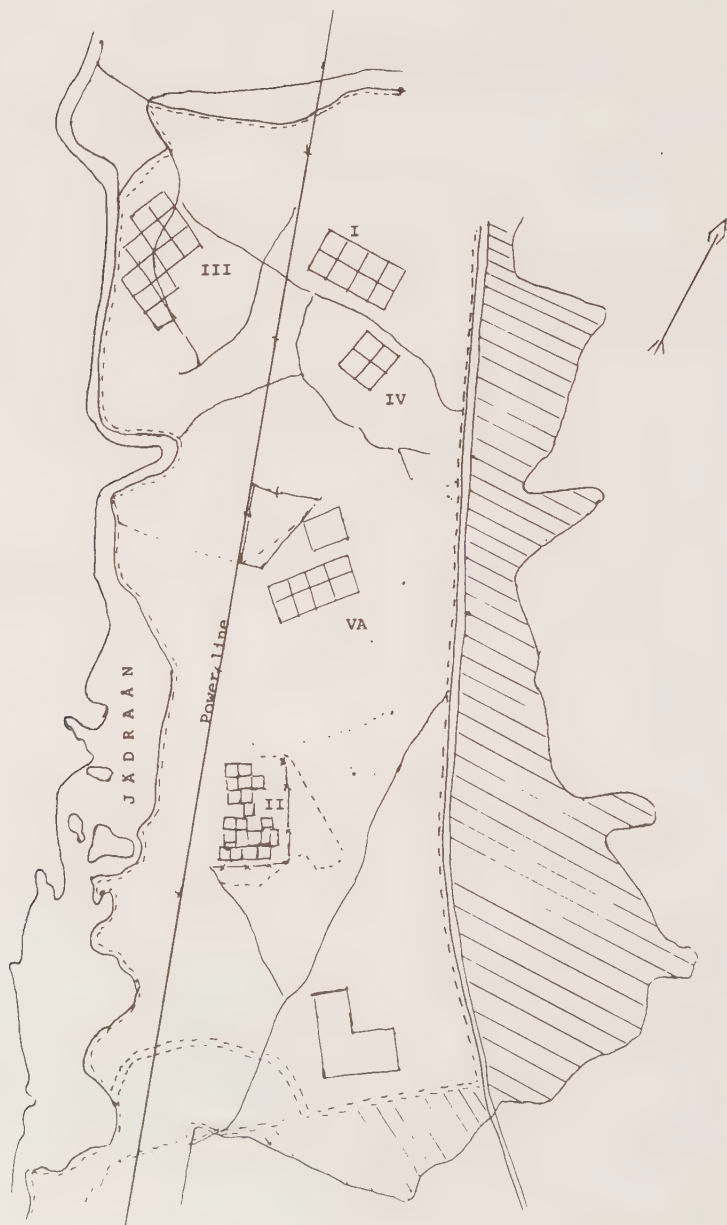
This site was selected for the present experiment because of the availability of the necessary base data collected within the Swedish Coniferous Forest Project. Fig. 29 shows the experimental site with the positions of the plots selected for tritium injections. The glaciofluvial deposit is underlain by leptite, porphyry and gneiss-granite bedrock. The soil profile is unusually dry. As a consequence, samples for tritium analysis could be obtained only from two plots (IH VA 1 and IH VA 2) out of the five selected ones.

The particle size distribution studies show that the profiles are mainly composed of silt, fine sand and medium sand with the fine sand as the dominant part. (Jansson, P.E., 1977). It is also observed that the dominance of fine sand increases with depth.

6.2 SKEDIGA

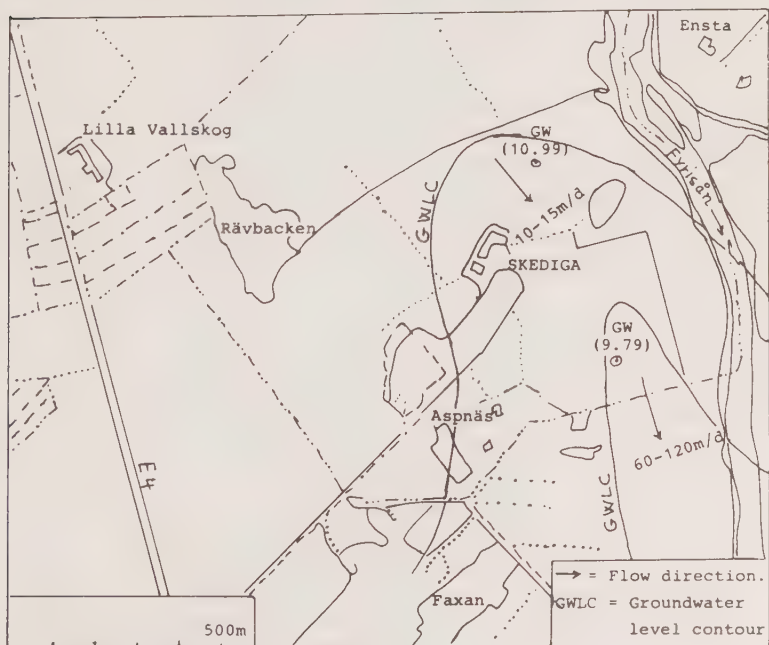
The topography of the bedrock of Uppsala area is characterised by its flatness and only disturbed by the occurrence of archaean rocks along the fault lines which run mainly in SE-NW direction. Along the fault line, the rocks are lifted upward on the eastern side and depressed on the western side thus forming a kind of valley which is filled with the post-glacial deposition. The Uppsala esker, on its southern part, follows the fault line while on the northern part it deviates from the fault line and occurs as rather isolated and spread hills (esker centers) in a NNW-SSW direction. The archaean formation is overlain by the post-glacial sediment (moraine on the Archaean outcrops, glacial and post-glacial clays on the plains and valleys). The clay layer can be

Fig. 29 A magnified version of the experimental site, Jädraås



as thick as 75 m with the eskers as the dominating glaciofluvial deposit in the region. As far as the hydrogeology of the area is concerned, the esker is the main groundwater storage formation. The groundwater level decreases from north to south reaching the level of Fyris river at Nedre Föret. The groundwater formation is also characterized by the occurrence of springs which provide much water for consumption (eg. S:t Eriks spring and the hospital spring). The groundwater discharges into river Fyris at Flottsund and feeds lake Ekoln. Most of the names of places mentioned above are outside the sketch of the experiment area shown. More than 70% of the water supply of Uppsala town comes from the esker formation. To refill the highly exploited groundwater storage, there is a big scheme of artificial recharge system at Tunåsen (Storvad). The

fig. 30 Groundwater levels and flow directions for Skediga area.



water from Fyris river is pumped up to the infiltration area situated about 2 km north of Tunåsen. At the experimental site (Skediga), the groundwater level lies 10 to 12 m below the surface. The groundwater level variations with the flow velocities are shown in figure 30. The soil profile studies show a layering of individual profiles and a big areal variation between the different profiles.

6.3 TÄRNSJÖ AREA

The glaciofluvial deposits making the esker formation of the area are underlain by the precambrian granites which are assumed to have a good hydraulic conductivity because of the fissured conditions. The water divide falls along the top of the eskers as shown in fig. 31. There are several springs with discharges varying between 30 to 100 l/s. Some examples of such springs are Östa spring (50 - 100 l/s) and Ulebo spring (30 - 50 l/s) (Norberg and Persson, 1974).

There are two aquifer systems in the area. The groundwater level of the deeper one is as deep as 100 m below the surface while the upper one can be 7 - 8 m above the deeper levels (Ibidem). The groundwater divide with the flow directions is shown in fig. 31. Using the base material obtained from the Swedish Geological Survey (SGU), the different esker materials can easily be localised for the selection of injection plots at the representative points.

Fig.31. Map showing the water divide and the direction of groundwater flow in Tärnsjö area.



7. MEASUREMENTS AND ANALYSIS

7.1 SOIL PROFILES

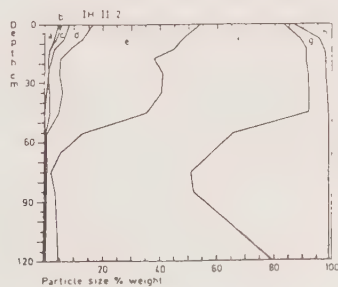
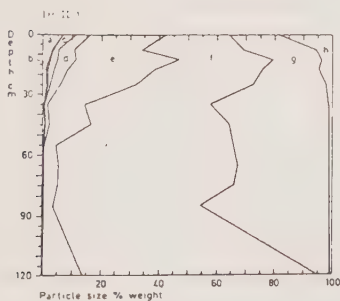
Soil profile studies are done for the three plots in Skediga. As for the two plots in Jädraås the necessary information is obtained from Per Erik Janssons work, 1977. No detailed profile studies are available for the Tärnsjö sites.

7.1.1 JÄDRAÅS

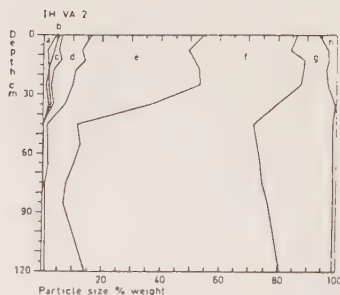
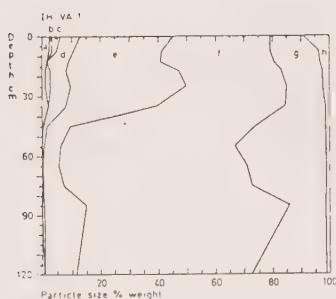
The study of the particle size analysis of the soil profiles in the Jädraås sites shows that the top soil layer down to 120 cm is relatively homogeneous. There is a slight dominance of finer material at the top 20 - 30 cm layers. Below 80 cm the finer material increases with depth. The other two plots are included to show the areal variability of the soil material. (see figs. 32 and 33)

Fig. 32 and 33 Soil profiles at Jädraås. (From Jansson, P.E., 1977)

32



33



Legend: a = < .002
 b = .002 - .006
 c = .006 - .02
 d = .02 - .06
 e = .06 - .2
 f = .2 - .6
 g = > .6
 h = Ignition loss
 (Organic material)

7.1.2 SKEDIGA

In order to perform soil profile analysis for Skediga plots, 10 cm cylinders were used to take soil samples at 10 cm intervals in the vertical direction. The grain size distributions for the plots B, C and D are shown in figure 34. All the three profiles show that the soil material is layered. In plot B the coarser material dominates down to 35 cm below which fine and medium sand dominate to a depth of about 75 cm. Below 75 cm depth the coarser material again dominates. Plot C shows a fine-coarse-fine-coarse layering at 0 - 50 cm, 50 - 65 cm, 65 - 85 cm and 85 - 110 cm respectively. Plot D which lies at the top of the esker is dominated by coarse sand and gravel down to 90 cm depth. The effects of these stratifications are discussed in the theory, section 3 and in the results and discussion in section 8. Table 1 and figure 34 show the results of the grain size analysis.

7.2 EVAPOTRANSPIRATION AND PRECIPITATION DATA

7.2.1 JÄDRAÅS

The accumulated evapotranspiration and precipitation values for Jädraås area and for the observation period are given in table 2. The evapotranspiration values are reduced to the actual losses using the soil moisture factor as shown in equation (53).

Table 2 Evapotranspiration and precipitation data for Jädraås.

Date	Accum. Prec.	Accum. Evapotr.
760922	94	16
760929	96	20
761007	108	25
761009	108	27
761112	235	40
761113	235	40
770604	636	120
770615	655	150
770618	655	158
770627	691	174
770707	732	195

7.2.2 SKEDIGA

The potential evapotranspiration for Skediga area is estimated by means of the Penman formula with the meteorological data from the nearby station, F16. In fig.35 and table 3, evapotranspiration values by the Penman and Penman-Monteith formulas are compared with the corresponding

precipitation values. The temperature variations are also included. The Penman-Monteith values are much lower than the Penman values and are supposed to be estimates of the actual evapotranspiration. The last column in table 3 shows a deficit of moisture all through the period when the Penamn estimate is used and an excess of moisture all through the period when the Penman-Monteith estimate is used. Fig.35 shows also the the cyclic variation of the three variables evapotranspiration, precipitation and temperature are similar for the four years. It can be seen that 1976 was a relatively dry year. The daily values with the superimposed monthly sums are shown in figs.36A - 36D

7.2.3. TÄRNSJÖ

For Tärn sjö area where no meteorological data are available, data from Ljusbäck was used. The potential evapotranspiration is estimated by means of the Skediga data.

Fig.35 A graph showing the relative variations of precipitation, evapotranspiration and temperature for the observation period (F16 data).

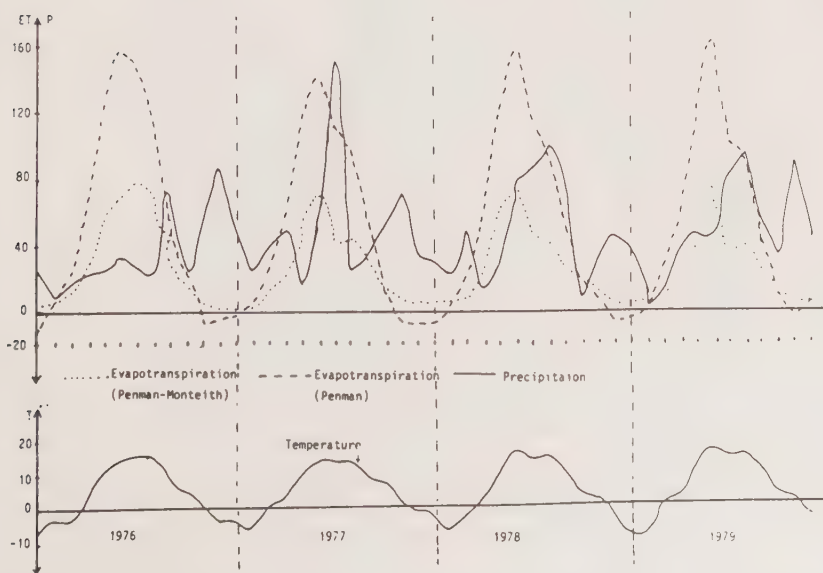
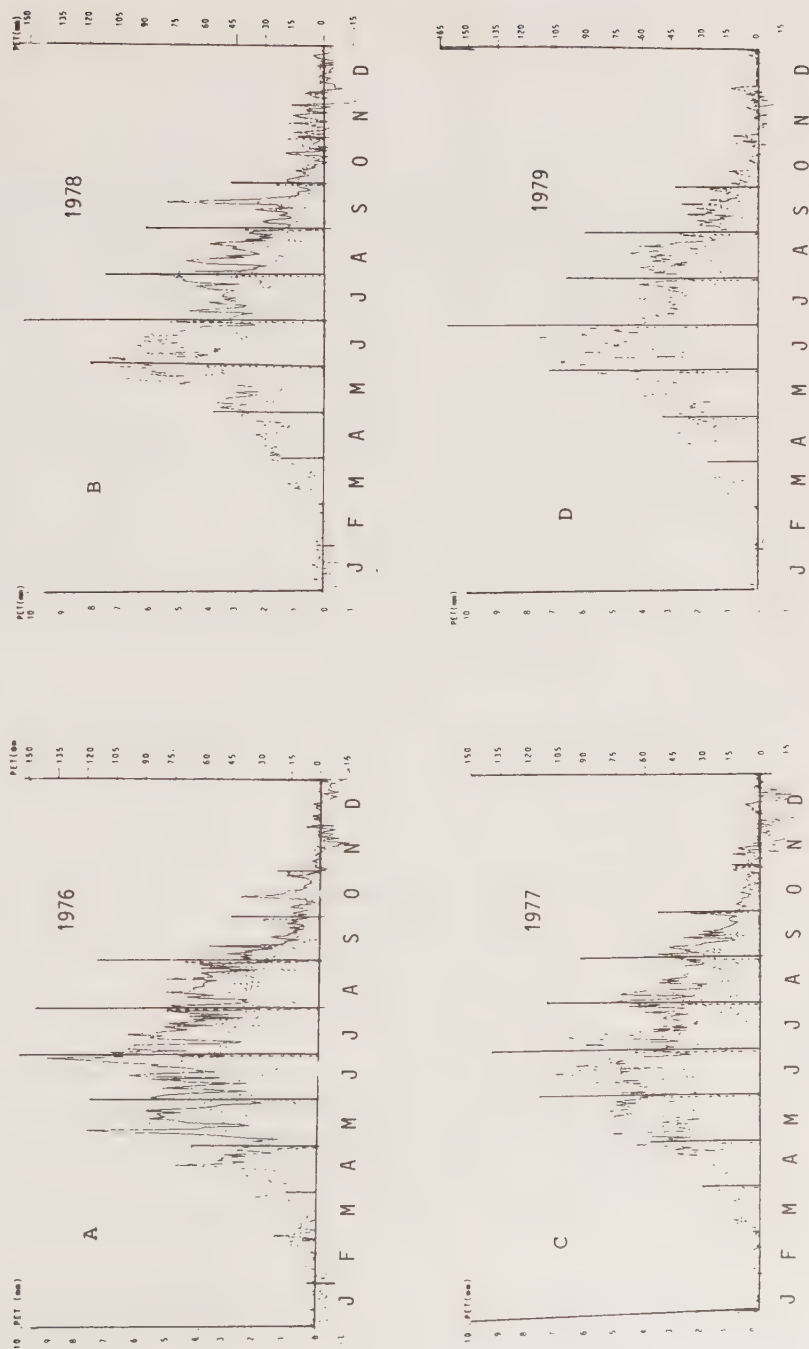


TABLE 3 MONTHLY VALUES FOR PRECIPITATION, EVAPORATION AND TEMPERATURE AT A METEOROLOGICAL STATION (F 16) NEAR SKEDIGA

No.		P yr												P-E	
		Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec		
1 1976	T	- 6,5	- 2,5	- 3,3	3,7	12,8	14,2	16,4	15,7	8,4	5,0	1,4	- 4,0		
	P	25,9	9,7	17,7	23,1	25,5	32,1	27,0	22,5	73,3	20,10	54,9	85,1	416,9	
	E	-11,983	6,012	21,277	67,432	119,269	156,58	147,446	116,19	45,88	22,03	- 7,54	- 4	677,8	-260,9
	E _m	3,1	5,4	9,6	24,2	55,6	69,0	77,6	69,6	28,8	16,7	4,7	1,0	365,3	51,6
2 1977	T	- 2,9	- 5,9	0,4	2,3	9,7	14,3	13,7	13,8	9,0	7,1	2,2	- 1,1		
	P	48,0	23,7	39,6	49,2	14,8	55,1	152,6	23,7	37,0	49,5	70,4	30,3	593,9	
	E _p	- 0,804	3,16	27,46	57,9	116,1	140,8	112,5	95,223	53,618	13,0	- 3,01	- 6,661	609,3	- 15,4
	E _m	2,7	2,2	13,0	19,6	53,0	66,9	41,4	44,1	30,3	13,0	7,0	4,9	298,1	295,8
3 1978	T	- 2,1	- 7,0	- 1,6	2,2	9,6	15,0	14,4	14,0	9,0	5,5	4,0	- 9,3		
	P	30,7	21,8	48,0	13,9	27,75	77,6	85,1	99,9	74,9	7,6	32,1	44,4	532,9	
	E _p	- 6,304	2,136	22,207	57,773	121,557	156,687	113,845	92,201	48,477	13,512	6,96	- 4,353	624,8	- 91,9
	E _m	5,0	3,9	6,9	22,9	61,5	71,9	45,5	40,7	25,1	18,9	13,3	3,6	317,1	215,8
4 1979	T	- 8,1	- 8,2	- 0,6	3,2	10,7	16,6	14,5	15,0	10,8	4,3	1,9	- 3,4		
	P	40,0	7,6	12,7	37,4	44,7	43,5	81,2	95,0	50,9	32,1	92,9	43,59	581,6	
	E _p	- 3,292	1,037	26,247	49,678	109,968	164,0	100,95	91,615	44,265	7,803	-0,43	6,98	585,0	- 3,4
	E _m	4,2	4,5	9,5	18,3	40,9	72,7	34,2	38,6	24,7	11,0	5,3	4,0	267,9	313,7

Legend: T = temperature E_p = evapotranspiration (Penman)
P = precipitation E_m = evapotranspiration (Penman-Monteith)

Fig. 36 The potential evapotranspiration, PET, at F16 near Skediga for the different periods indicated and as estimated by the Penman formula (solid line) and by the Penman-Monteith formula (broken line). The vertical lines are the monthly sums (Penman).



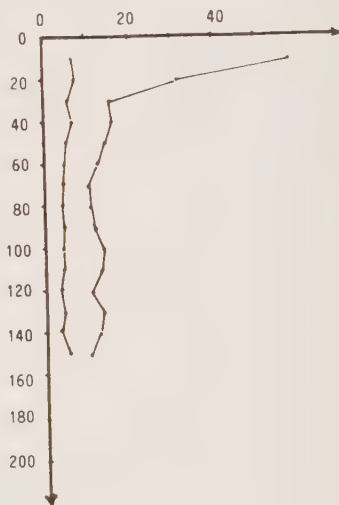
7.3 SOIL MOISTURE

Soil moisture data is procured for all the profiles in Skediga. For Jädraås plots which lie close together, soil moisture data from one access tube is used. In Tärnsjö area, soil moisture data is available only for one plot, plot C. There lacks access tubes at the other plots.

7.3.1 JÄDRAÅS

The soil moisture conditions of this area were thoroughly studied by P. E. Jansson (1977). Fig. 37 shows the distribution of the maximum and the minimum moisture contents along the profile during the observation period. The figure clearly indicates that the top 30 cm of the soil profile has the biggest moisture holding capacity. Below this level, up to 150 cm depth, the profile looks to be relatively homogeneous but a small disturbance at 70 and 130 cm levels. The higher water holding capacity near the surface depends upon the dominance of finer material. The soil moisture contents for the whole period and for the different observation periods are shown in tables 4, 5 and 6 (pp. 121 - 122). The values in tables 5 and 6 are used in the estimation of recharge by the tritium-tagging method the results of which are discussed in the next section.

Fig. 37 A graph of maximum and minimum moisture content.



7.3.2 SKEDIGA

The neutron soil moisture measurements for Skediga stations were done at 10 cm intervals up to 40 cm depth and at 15 cm intervals below this depth. This is done so in order to make the neutron measurements correspond to the tritium sampling depth intervals. The count rates in cpm are changed into soil moisture content by volume percent using the calibration equation (47). The results for the plots A - B are shown in tables 7A - 7D (pp. 123 - 125). These values are used in the estimation of ground water recharge by the tritium-tagging method.

The average soil moisture contents for the different periods and depths are presented in table 8 (pp. 126). Tables 9A - 9D (pp. 127 - 128) show the summations of soil moisture contents in mm for the depths 0 - 50 cm, 0 - 100 cm, 0 - 200 cm, 50 - 100 cm and 10 - 200 cm and their variations between the different sampling occasions and the depth intervals. The soil moisture summations for the depth intervals of 0 - 100 cm and 100 - 200 cm are also shown diagrammatically in figs. 38A and 38B. It can be observed that in the first 100 cm layer of the soil profile, more moisture is stored in plot B than plot C (fig. 38A) while below this level plot C contains markedly more moisture than plot B. Plot D which has a profile with very coarse material contains the least moisture as compared to the other two. This difference in moisture containing capacity depends on the different layering sequences. The plot of the maximum and minimum soil moisture contents (fig. 39) predicts the same soil properties as above.

7.3.3 TÄRNSJÖ

In Tärnsjö area, soil moisture measurements are available only for station C. A groundwater observation tube is used for the Danish-type neutron moisture probe. Soil moisture contents by volume percent for the levels shown and for the period 1977 to 1979 are shown in table 11 (pp. 131)

Fig. 38A Soil moisture variations with time in the profile at 0 - 100 cm depth interval, Skediga.

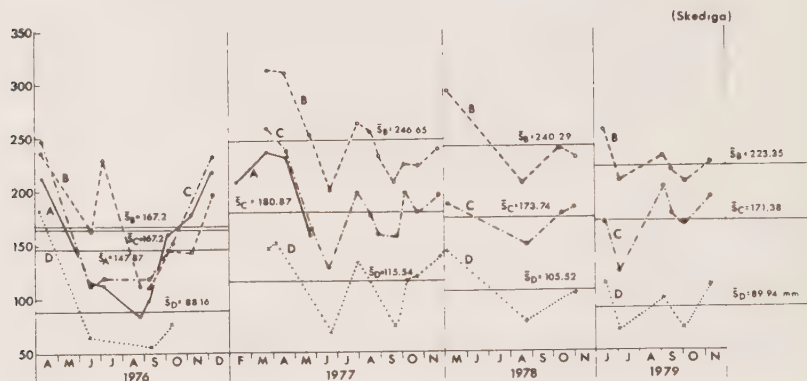


Fig. 38B Soil moisture variations with time in the profile at 100 - 200 cm depth interval, Skediga.

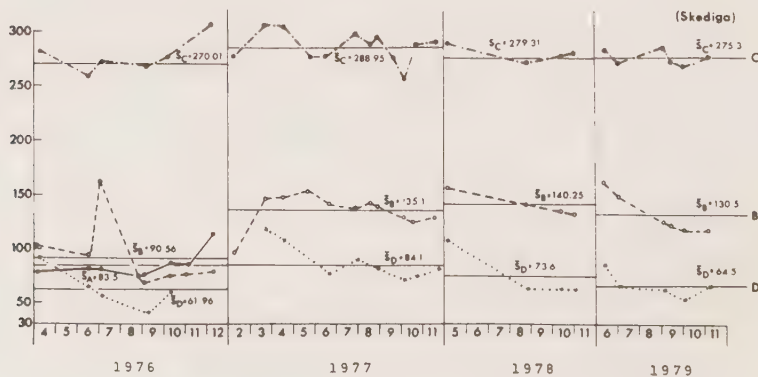
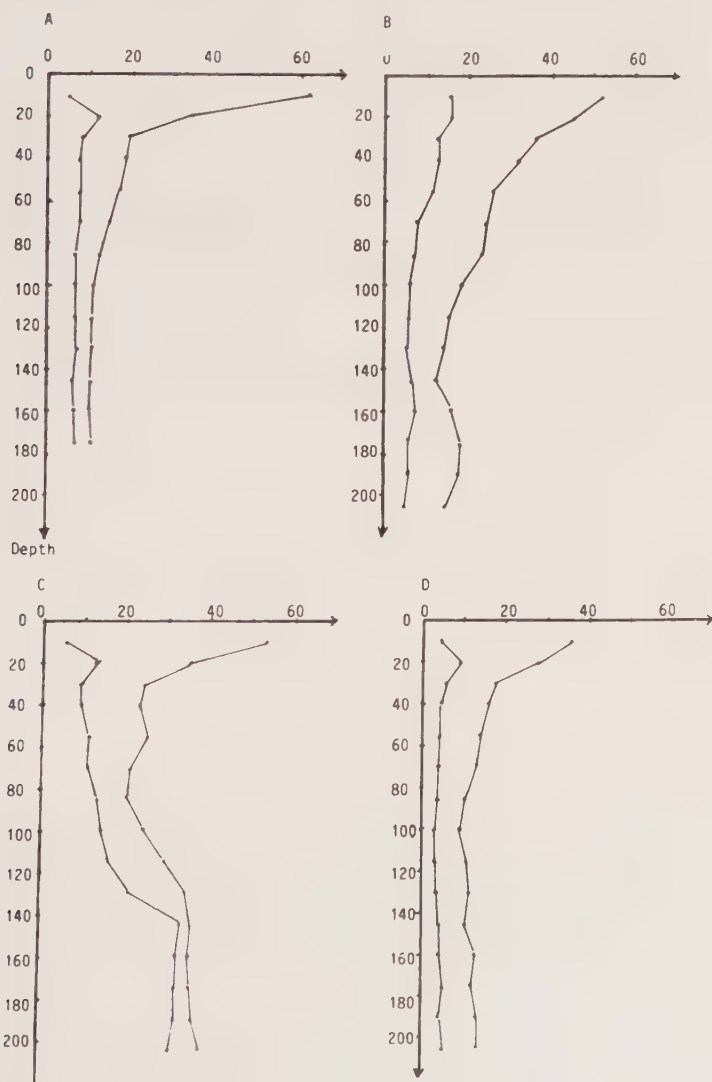


Fig. 39 A graph of the maximum and minimum soil moisture content
Skediga



Summations of the total soil moisture contents for the depth intervals 0 - 50, 50 - 100, 0 - 100, 100 - 200 and 100 - 200 cm are presented in figs. 41A - C. The graphs for 1978 and 1979 clearly show that maximum storage occurs during the period between March and April. The diagram of the maximum and minimum soil moisture content along the profile as shown in fig. 40 indicates the layering of the soil profile into sections of 0 - 50, 50 - 150 and 150 - 200 cm.

The soil moisture values calculated for the observation period are shown in table 12 (pp. 132). One can, at a glance, observe that the moisture values are highest in 1977 and least in 1979 and correspond to the average monthly precipitation which is highest in 1977 (15.66 mm) and least in 1979 (11.3 mm). See figs. 41A and 41B.

Fig. 40 A curve of the maximum and minimum soil moisture contents for Tärnsjö site.



Fig. 41A Soil moisture (plot C), precipitation and temperature data for Tärnsjö, 1977

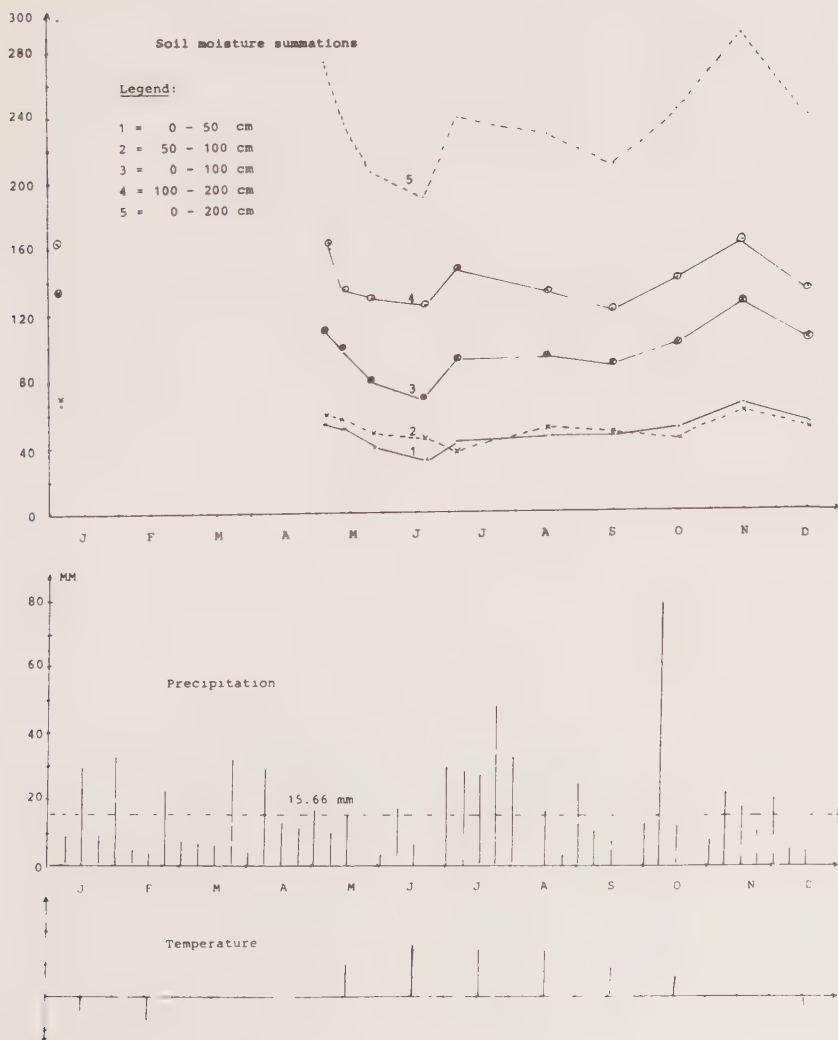


Fig. 41B Soil moisture (plot C), precipitation and temperature data for Tärnsjö, 1978. (1,2,3,4 as shown in fig. 41A)

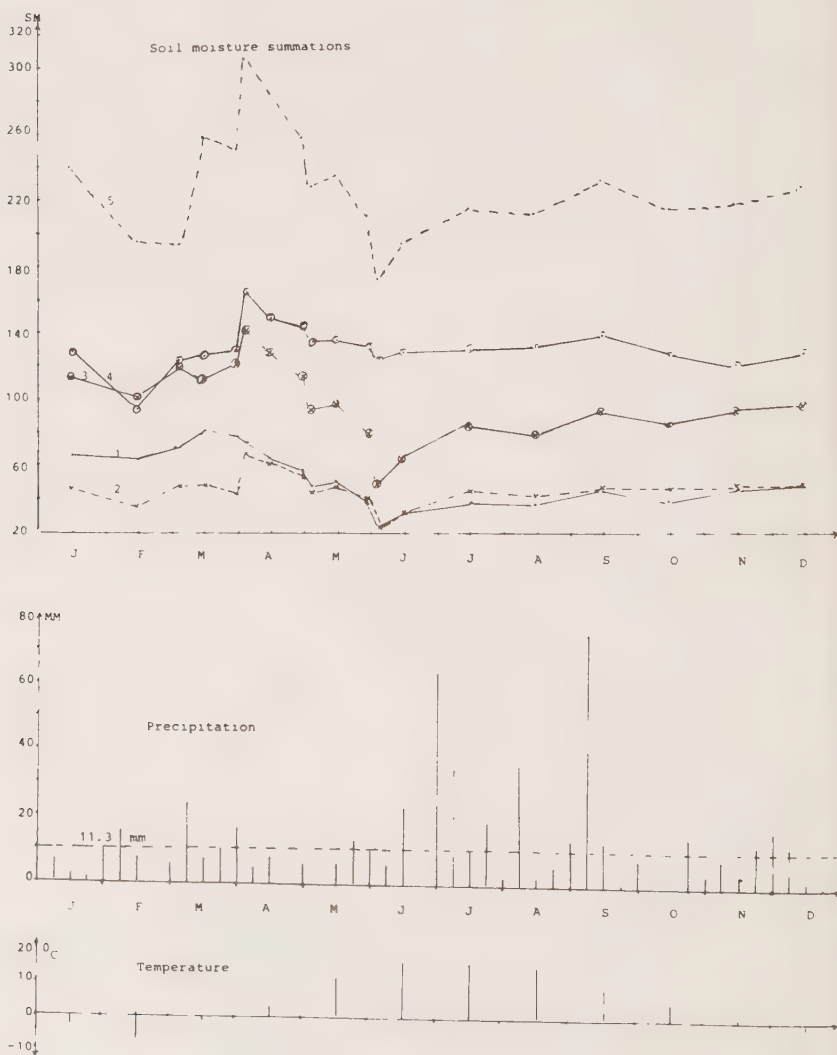
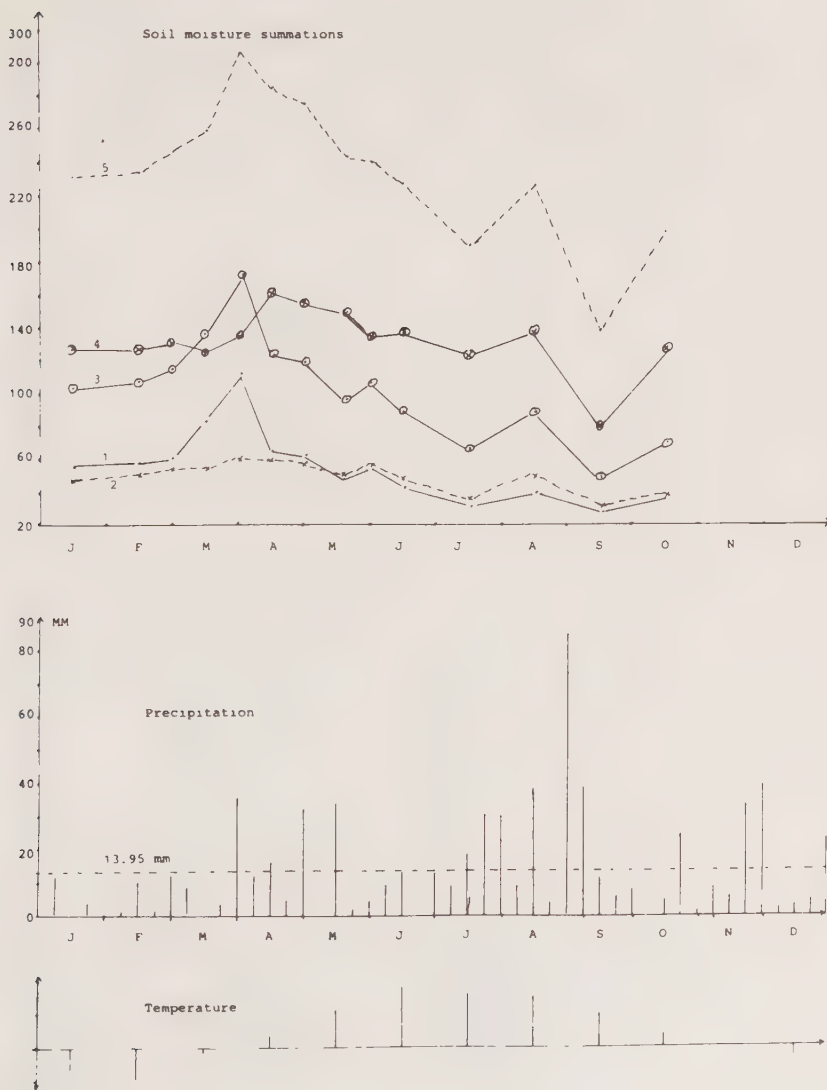


Fig. 41C Soil moisture (plot C), precipitation and temperature data for Tärnsjö, 1979. (1,2,3,4 as shown in fig. 41A)



7.4 TRITIUM PROFILES

The tritium profiles are drawn with the tritium contents in cpm (counts per minute) on the horizontal axis and with the depth scale in cm along the vertical. Since it is the displacement of the peak concentration one is interested in, it was not necessary to calculate the tritium units (TU). When the peak concentration is not so obvious, the center of gravity of the tritium distribution (CG) was used to estimate the tracer displacement.

7.4.1 JÄDRAÅS

At Jädraås, where extensive field experiments were performed, tritium profiles for the two stations IH VA 1 and IH VA 2 are shown in figures 42 and 43. The problem of obtaining enough water samples for tritium analysis was specially big in this area so that pumping had to be continued for longer periods than usual. This prolonged pumping may cause an external pressure thus artificially increasing the vertical flow velocity. The 1976 profiles, especially after the 22nd of June, for station IH VA 1, are difficult to interpret partly because the profiles are not well developed and partly because of the missing samples at several levels. The latter is true for both the stations and for the whole period. The 1977 profiles for the same plot, with pronounced vertical distribution, are also less developed but could be used for estimating recharge.

The 1976 profiles for IH VA 2 show an upward movement of the tracer but this may also be a second peak. Anyhow, the average displacement is used to calculate recharge estimates. The 1977 profiles for the same plot show a slow vertical movement accompanied by a pronounced vertical distribution of the tracer. The peak is, however, well developed. The average vertical velocities for the different periods are calculated and shown in table 13.

Fig. 42 Tritium profiles for station IH VA-1, Jädraås.

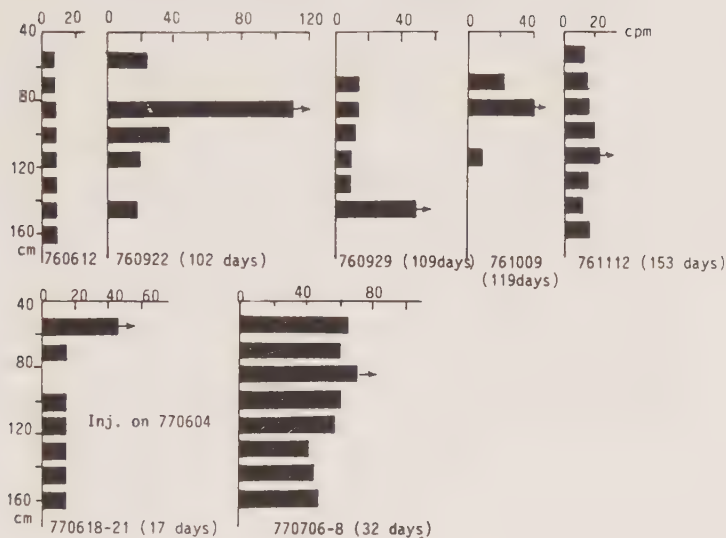


Fig. 43 Tritium profiles for station IH VA-2, Jädraås

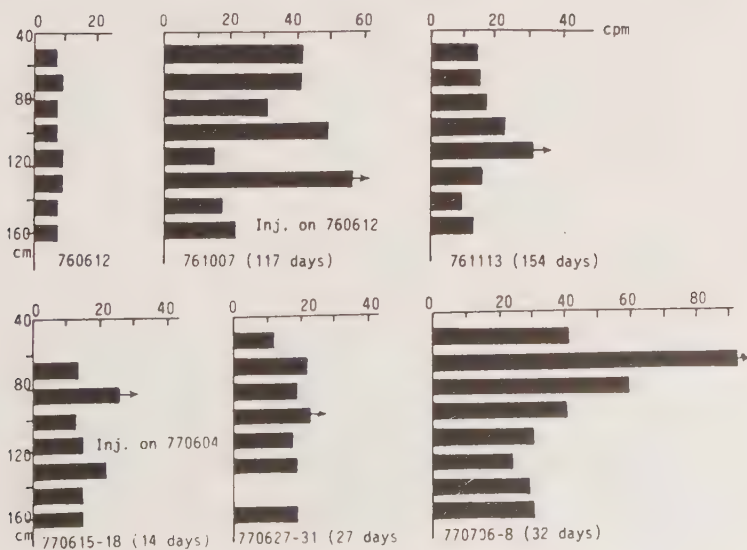


Table 13 The peak displacement rates in cm/day, Jädraås

Stn.	Period	t (day)	Displacement (cm)	Depth (cm)	dx/dt (cm/day)	$\bar{v}$ (cm/day)
IH VA 1	760612-0929	109	105	145	0.96	1.19
	770604-0706	32	45	85	1.41	
IH VA 2	760612-1007	117	90	130	0.77	0.855
	770604-0706	32	30	70	0.94	

$\bar{v}_t$ (average for the whole period and area) 1.03

7.4.2 SKEDIGA

PLOT A: The 1976 profile for plot A (which was abandoned in the beginning of 1977 because of a house building) shows a distinct peak at 85 cm depth after 101 days and at 130 cm depth after 127 days. The broadened shape and position of the second peak indicates increased diffusion process. The average rate of vertical moisture movement during the whole period was 11.3 mm per day and the velocity of water molecules was 18 mm per day. See fig. 44.

PLOT B: For plot B (fig.45), the 1976 tritium peak is poorly developed and the peak lies far below the center of gravity. However, the displacement of the center of gravity gives better recharge values. The profiles for 1977 and 1979 show no peak displacement below 55 cm depth. From grain size distribution studies, this level is dominated by fine sand with 20% clay content which reduces the hydraulic conductivity. The downward movement is thus retarded. See fig.45 The 1978 profile shows no peak and the CG moves up and down. Therefore this profile can not be used to calculate recharge.

PLOT C: The tritium profiles for this plot show that the tritiated water is highly distributed along the soil profile. We can see for example the 1976 tritium profile is specially flattened and wide. This is due to the fact that 1976 was a dry year and it seems that under drier conditions the dispersion effect is greater. It is also observed in fig. 45 that the ^3H -peak is sustained at 90 cm depth. From the soil profile studies, at this level the fine sand fraction dominates thus reducing the vertical displacement rate. The 1978 tritium profiles are not well developed.

PLOT D: This plot is situated on the top and coarser part of the esker material. As a result it has been difficult to extract water from all levels for tritium analysis using the present sampling method. Although the material is very coarse, the relative layering of the profile with a finer-over-coarser sequence (40 to 65 cm) makes the tritium peak to be sustained for a long time. See fig. 47 The 1976 peak is unreliable because water samples are missing for several levels. The same is true for the 1978 profile. The 1977 and 1979 tritium peaks are used for recharge calculations. The profile for 1976 shows an upward movement of tritium-tagged water due to the dry conditions and increased evapotranspiration loss from the soil profile.

The rates of moisture movement along the profiles and for the different periods are shown in table 14

Fig . 44 Tritium profiles for Shediga. plot A

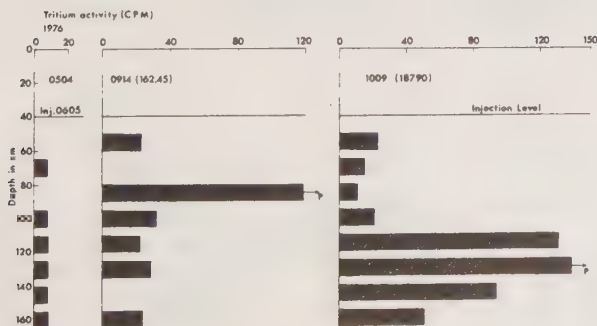


Fig. 45 Tritium profiles for Skediga, plot B

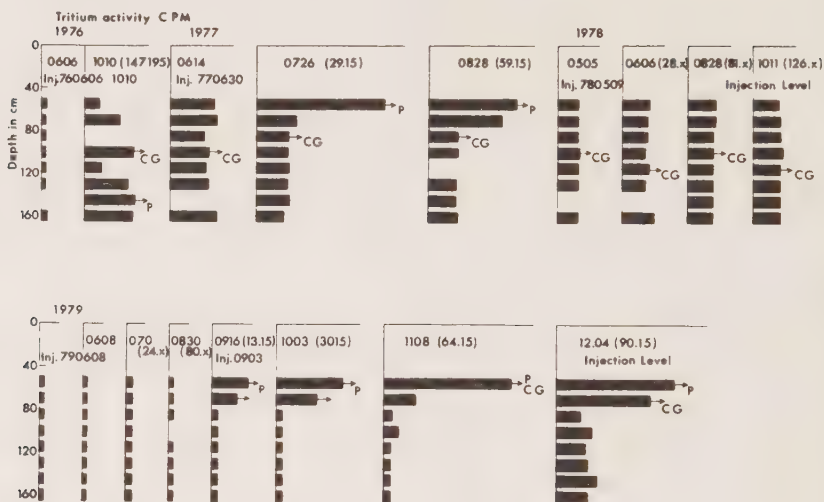


Fig. 46 Tritium profiles for Skediga, plot C

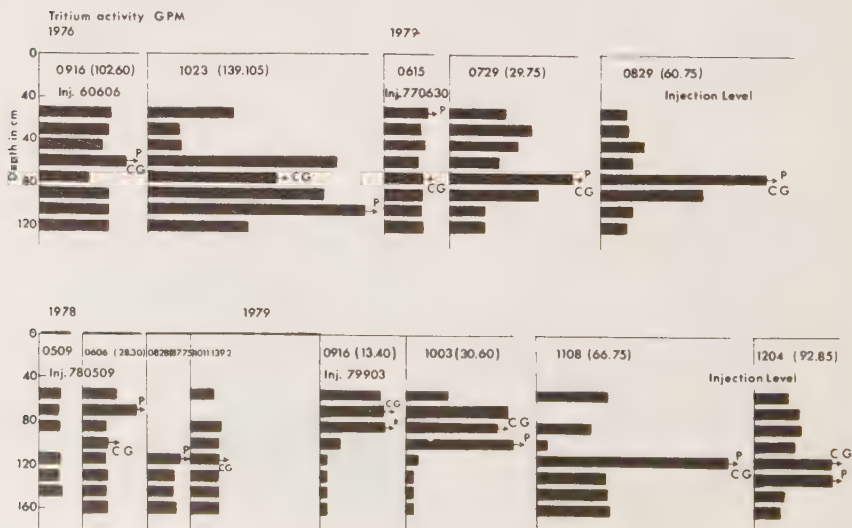


Fig. 47 Tritium profiles for Skediga, plot D

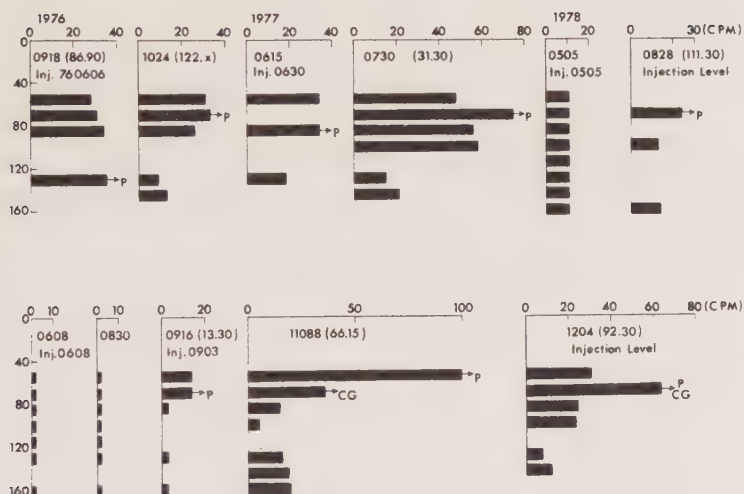


Table 14 Percolation rates for Skediga plots

Stn.	Period	t (days)	Depth (cm)	Displa- cement	dx/dt (cm/day)	$\bar{v}$ (cm/day)
A	760605-0914	101	85	45	0.45	1.13
	760605-1009	126	130	90	0.71	
B	760606-1010	126	100	60	0.48	1.19
C	760606-0916	102	100	60	0.59	0.50
	760606-1023	139	115	75	0.76	
	770630-0729	29	115	75	2.60	1.95
	770630-0829	60	115	75	1.30	
	780509-0606	28	70	30	1.10	1.10
	790903-1204	92	115	75	0.82	0.82
D	760606-0918	104	130	90	0.87	1.35
	760918-1024	37	70	-60	1.62	
	760606-1024	141	70	30	0.21	
	770630-0730	31	70	30	0.97	0.97
	780505-0828	115	70	30	0.26	0.26
	790903-1204	92	70	30	0.33	0.33

$\bar{v}_2$ (average velocity for the whole period and area) 0.96

7.4.3. TÄRNSJÖ

The tritium profiles for the five plots in Tärnsjö area are presented in figs. 48 to 52. Even here data for some levels is missing because of the behaviour of the porous cells in esker materials where either the contact between the cells and the soil material is poor or the soil suction force is so great that no water is sucked into the porous cells.

PLOT A: From the general geological considerations by (Norberg and Persson, 1974) this plot should lie in a sandy area but observations during the installation of the sampling tubes, have shown the occurrence of coarser material just below 100 cm depth. The 1978 and 1979 profiles have, to a certain extent, the shape of a Gaussian distribution. The 1977 profile on the other hand, shows the dominance of dispersion for the shape is broadened and more distributed along the soil profile.

PLOT B: This plot is situated in a very coarse, gravelly and stony area that a symmetrical installation of sampling tubes was very difficult. One is forced to try several times at different spots to come to a certain depth. At some levels it is difficult to suck moisture. This may be due to the occurrence of bigger stones making the contact of the cells with the soil very poor. The 1977 and 1979 profiles show very little peak displacement as compared to the 1978 profile possibly because in the former case samplings were done during the later part of the summer period when vertical movement downwards is either minimal or reversed due to evapotranspiration where as in the latter case samplings were done just after the snow melt when moisture movement is primarily vertically downwards.

PLOT C: The profiles for plot C show that the soil moisture is sustained near the injection level during all the observation periods. This may be due to the layering of the soil, most probably a fine-over-coarse sequence because the diagram of the maximum and minimum soil moisture content (fig. 41) shows a high moisture capacity down to 60 cm depth and a deeper layer of relatively lower moisture capacity. The disappearance of the 1978 peak may be due to the fact that when the suction force

at the upper boundary equals or exceeds the suction force at the lower boundary, there occurs a fast vertical drainage in the profile. This phenomena is discussed in detail in section 3.

PLOT D: Plots B and D are located in a similar formation, very coarse, stony and gravelly material. The profiles of the two plots show similar behaviour, flattened and highly distributed tracer profiles which are typical of dispersion dominated flow in a porous media.

PLOT E: The 1977 profile shows the distribution of tritiated water injected in the beginning of August. This is the late summer period when when very little downward moisture movement takes place. Hence the tritium peak seems to be stationary at the 50 cm depth. In 1978 tritiated water injected in the beginning of may (ie. just after snow melt) is followed. This profile shows distinctly a vertical moisture movement. The 1979 injection was done in the beginning of june and subsequent samplings show no vertical movement probably because the snow melt input had already passed the injection depth. This implies a fast vertical drainage after snow melt-input. The average particle velocities for the observation periods and for the different plots are shown in table 15.

Fig. 15 Percolation rates for the different plots in Tärnsjö

Stn.	Period	t (days)	Depth (cm)	Displa- cement	dx/dt (cm/day)	$\bar{v}_1$
A	770806-1009	64	115	75	1.17	1.31
	780512-0609	28	85	45	1.61	
	790904-1109	66	115	75	1.14	
B	770806-1007	62	70	30	0.48	1.02
	780512-0609	28	100	60	2.12	
	790904-1109	66	70	30	0.46	
C	770806-0903	28	75	35	1.25	0.70
	780512-0609	28	55	15	0.54	
	790904-1109	66	60	20	0.30	
D	770807-1009	63	85	45	0.71	0.80
	780512-0905	115	160	125	1.09	
	790904-1109	66	85	40	0.61	
E	770807-0911	25	55	15	0.54	0.47
	780512-0905	116	130	90	0.78	
	790904-1109	66	55	15	0.23	
$\bar{v}_2$						0.86

$\bar{v}$ = Average particle velocity for each plot and the whole period in cm/day.

$\bar{v}_2$ = Average particle velocity for all the plots during the whole period in cm per day.

Fig. 48 Tritium profiles for station A, Tärnsjö.

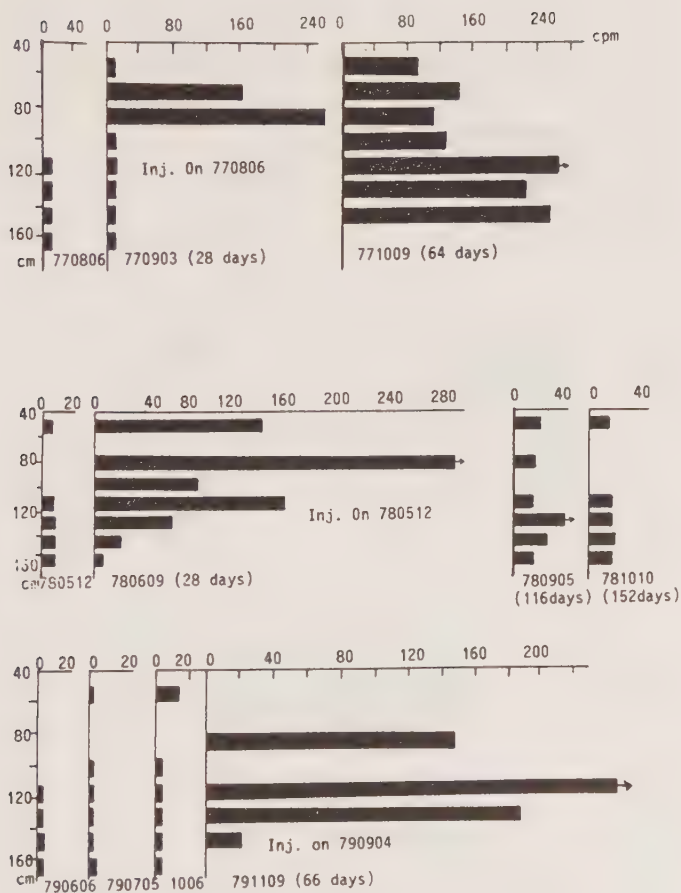


Fig. 49 Tritium profiles for station B, Tärnsjö

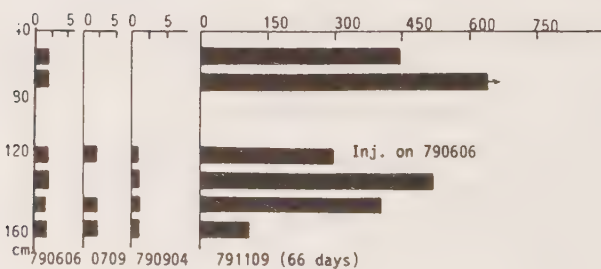
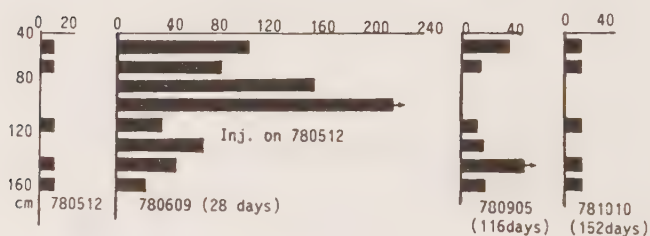
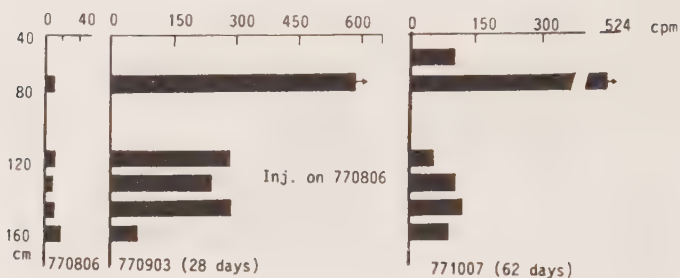


Fig. 50 Tritium profiles for station C, Tärnsjö.

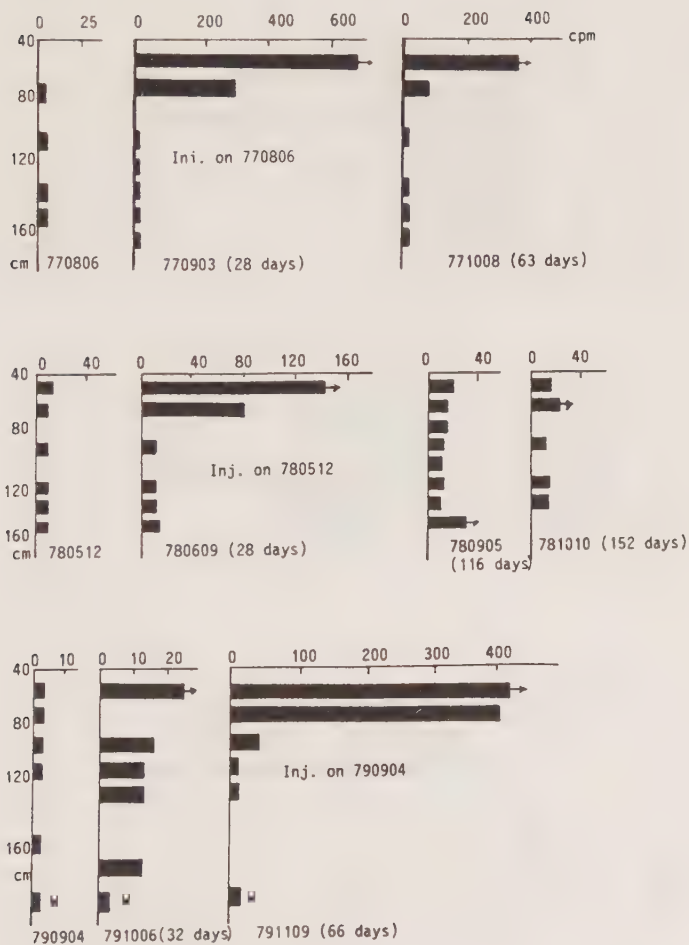


Fig. 51 Tritium profiles for station D, Tärnsjö.

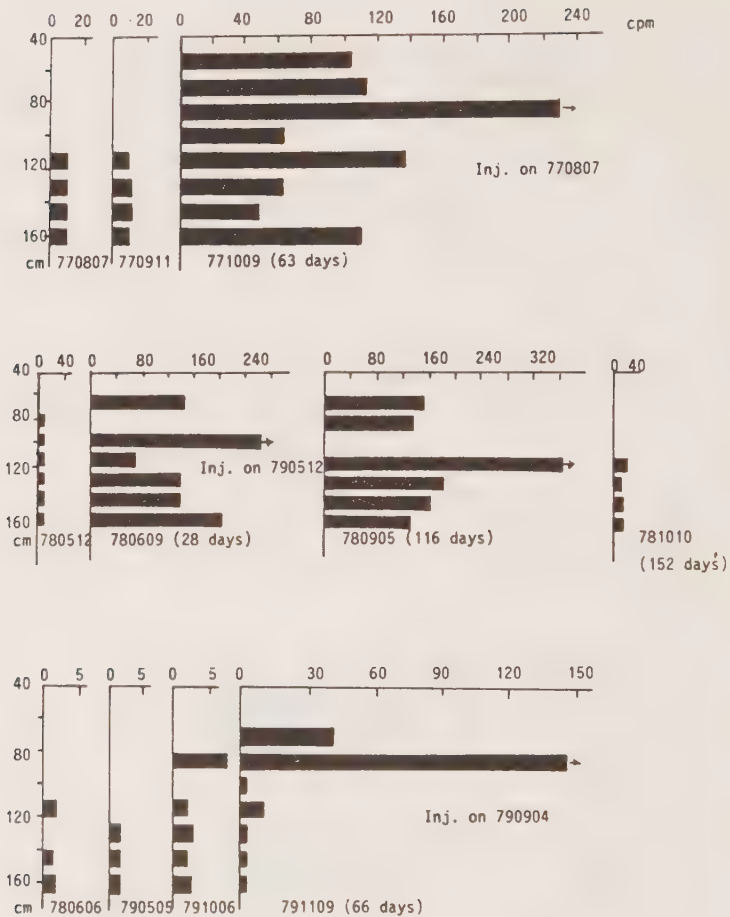
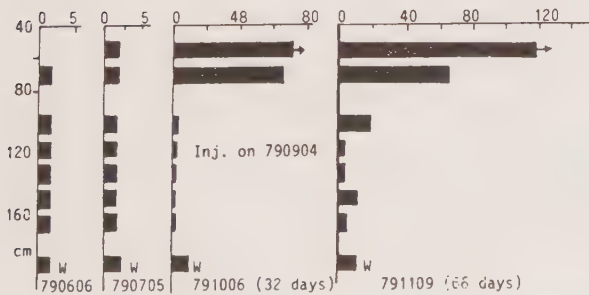
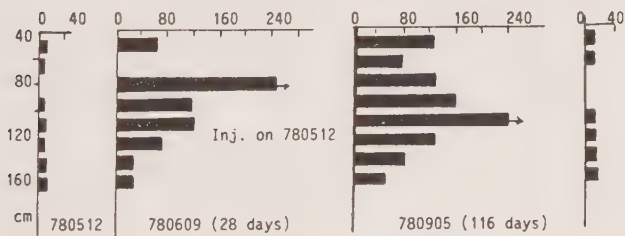
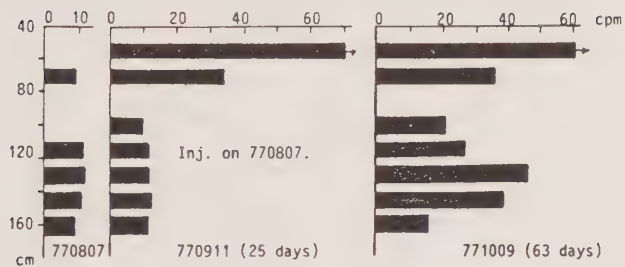


Fig.52 Tritium profiles for station E, Tärnsjö.



8. RESULTS

Percolation rates (particle velocities) for the different investigation areas and for the different observation periods are shown in tables 13, 14 and 15. Many important conclusions can be drawn from these studies.

Because of the large spatial heterogeneity of the soil material of an esker formation, the values of the percolation rates vary very much between the different plots in the same area. This large variation between the plots in the different investigation areas and the averages for the individual areas are summarised in the table below. Although the spread between the plots is large, the average values for each area do not differ so much. Of special interest is the almost perfect agreement

Table 16 Comparison of the percolation rates for the different sites

Site	Spread (mm/day)	Average (mm/day)
Jädraås	7.7 - 14.1	10.3
Skediga	2.1 - 26.0	9.6
Tärnsjö	2.3 - 21.2	8.6

between the Skediga and Tärnsjö values which can be accounted for by the fact that in both the investigation areas all the different esker materials were represented whereas for Jädraås the two plots were situated in quite similar material (sandy and fine-sandy material). This areal variation of esker soil material shows the importance of several such point experiments in the study of a recharge area. As the number of sampling points depends on the heterogeneity of the area, no general rule can be drawn.

The other important conclusion that can be drawn from the percolation studies is that the initial percolation rate just after the snow melt is very rapid and during the latter part of the summer period the vertical downward movement is retarded, stationary or reversed depending on the intensity of evapotranspiration loss. This is clearly exhibited by the

^3H -peaks which are sustained at the same level for a long time until they disappear by dilution or inter flow (lateral flow) in layered soil profiles. The 1979 profiles for plots B and D in Skediga, the 1977 profiles for plots B, C and E in Tärnsjö area and the 1976 profiles for plots IH VA 1 and IH VA 2 all show the same behaviour of initial fast vertical movement followed by a subsequent low, reversed or no vertical moisture movement. However, if the soil profile is layered with a fine-over-coarse sequence, vertical movement can take place even during the latter part of the summer period because in such layering sequences, the moisture in the upper boundary is temporarily stopped until the suction force in the upper boundary is reduced to or exceeds that in the lower boundary.

The third important fact that can be deduced from the tracer movement study is the effect of layering in the soil profile. The displacement of a tracer is strongly dependent on the soil layering as discussed in section 3. The phenomenon of lateral flow, double peaks and different penetration depths are all caused by soil layering. The dispersion/diffusion process dominant in layered and coarse materials is the other phenomena that must be accounted for in soil moisture movement studies.

The recharge estimates for the different sites and periods by using the tritium-tagging method and the water balance method are presented in tables 16 - 22.

The recharge values as calculated by the ^3H -tagging method for Skediga areas are shown in table 16. As seen in the table some of the profiles which are not well developed are not used to estimate recharge. The 1976 estimates of recharge are 95.5 mm, 91.2 mm, 125.2 mm and 66.6 mm for the plots A - D respectively. The estimates for plots A and B agree closely because these two plots are in the sandy area where the profiles are relatively similar. A higher estimate is obtained for plot C partly because this profile is layered and partly because the observation period is longer (139 days as compared to 126 days for the other two.) Plot D which is situated on the ridge of the esker very near to the sharply inclined surface underestimates recharge probably due to lateral flow.

Table 16 Recharge estimation by the tritium tagging method, Skediga.

Stn.	Period.	Depth interval, (cm)	Shift, x (in mm)	Time, t (in days)	dx/dt (mm/d)	CMC, (vol. %)	Rech. R (in mm)	Prec. (mm)	r=R/P
A	760505-								
	760914	40 - 85	450	101	4.50	10.82	60.2	121.2	0,5
	760605-								
	761009	40 - 130	900	126	7.10	10,61	95.5	163.6	0,58
B	760606-								
	761010	40 - 100	600	126	4.80	15.20	91.2	161.4	0,57
C	760606-								
	760916	40 - 100	600	102	5.90	11.91	71.5	151.5	0.47
	760606-								
	761023	40 - 115	750	139	7.60	16.69	125.2	163.2	0.77*
	770630-								
	770729	40 - 115	750	29	26.00	17.01	127.6	152.6	0.84
	770630-								
	770829	40 - 115	750	60	13.00	17.28	129.6	176.3	0.74
	780509-								
	780606	40 - 70	300	28	11.0	18.93	56.8	28.2	2.00*
	790903-								
	791204	40 -115	750	92	8.2	18.51	138.8	162.8	0.85
D	760606-								
	760918	40 - 130	900	104	8.70	7.40	66.6	151.9	0.44
	760918-								
	761024	130 - 70	-600	37	16.20	7.13	42.8	11.3	3.80
	760606-								
	761024	40 - 70	300	141	2.10	7.05	21.2	163.2	0.13
	770630-								
	770730	40 - 70	300	31	9.70	10.88	32.6	152.6	0.21
	780505-								
	780828	40 - 70	300	115	2.60	12.23	36.7	283.8	0.13
	790903-								
	791204	40 - 70	300	92	3.30	9.73	29.2	162.8	0.16

Table 17. Groundwater recharge estimated by the water balance method, Skediga.

Stn.	Period	Number of days	AET (mm)	dS (mm)	P (mm)	R (mm)	r
A	760408-1206	242	285.5	4.32	358.0	68.18	0.20
	770208-0523	103	170.9	-43.90	125.9	73.40	0.58
B	760407-1206	243	344.9	-37.53	358.0	51.13	0.14
	770321-1126	250	439.0	-75.36	497.4	133.76	0.27
	780505-1101	180	298.5	-62.46	342.0	105.96	0.31
	790612-1108	149	128.7	-31.11	229.0	131.41	0.57
C	760408-1206	242	251.5	-15.41	317.0	80.91	0.26
	770208-1121	286	399.8	-2.04	497.0	99.24	0.20
	780507-1101	178	244.8	-5.64	342.0	102.84	0.30
	790612-1108	149	122.2	22.38	229.0	84.42	0.38
D	760408-1010	185	205.2	-106.4	181.0	82.20	0.45
	770322-1121	244	364.5	7.06	444.6	75.06	0.17
	780505-1101	180	269.0	-39.21	342.0	112.21	0.33
	790612-1108	149	174.2	-2.35	235.0	63.15	0.27

Table 18. Recarge estimate by the water balance method for shorter periods, Skediga.

Plot	Period	Number of days	ET (mm)	dS (mm)	P (mm)	R (mm)	r
A	760605-1009	126	128.1	-52.8	163.6	88.3	0.54
B	760606-1010	126	156.0	-93.0	161.4	96.4	0.60
C	760606-1023	139	169.5	-95.9	163.2	89.7	0.55
	770630-0829	60	70.0	-4.5	176.3	110.8	0.63
	780509-0606	28	40.0	-38.1	28.2	26.3	0.93
	790903-1204	92	10.3	68.9	192.8	162.8	0.51
D	760606-1024	141	183.3	-106.4	163.2	86.3	0.53
	770630-0730	31	74.6	-19.6	152.6	97.7	0.65
	780505-0828	115	169.6	-59.6	283.8	173.9	0.61
	790903-1204	92	34.8	42.5	162.8	85.5	0.53

The 1978 estimates for plots C and D (129.9 mm and 32.6 mm respectively) are difficult to compare directly because of the different observation periods (60 days for C and 31 days for D). But the comparison of the recharge values as a fraction of precipitation during the period (r) shows the same tendency of giving very high recharge estimates for plot C and much lower recharge estimates for plot D. The extremely high recharge value for plot C and for the period 780509 to 780606 (28 days) indicates that very short observation periods give erroneous results.

Tables 17 and 18 show the recharge estimates obtained by the water balance method using equation (1). Table 18 contains recharge values corresponding to the observation periods for the tritium-tagging method. In general the water balance method seems to give higher recharge values than the tritium-tagging method. A close comparison of the two methods is done further down.

The groundwater recharge values for Tärnsjö area and for the short periods indicated are given in tables 19 and 20. As in the case of Skediga, because of the differing length of observation periods, direct comparison of recharge values can not be made. Therefore the quotient of recharge to precipitation during the same period (r) is used for comparative comments.

As seen from the results, the recharge values for plot C are very low as compared to the other plots. Also the 1978 estimate for plot B gives extremely high recharge value, 1.27 times the precipitation value for the period, because of the relatively short observation period (28 days). If we exclude these few anomalies, about 90% of the recharge values lie between 13% to 45% of the precipitation values during the same period. This is in the range of values as obtained by other workers like Sukhija, 1972; Athavale et al. 1978 and Zimmermann et al., 1966, 1967. The estimates of recharge by the water balance method for this area are shown in table 21.

Table 19 - A table for the estimation of recharge by the ^3H -tagging method, Tärnsjö.

Stn. Period	Depth interval, (cm)	Shift, x (in mm)	Time, t. (in days)	dx/dt (mm/d)	CMC, ϕ (vol. %)	Rech R 0. x (mm)	Prec. (mm)	r=R/p
A	770806- 770903	40 - 85	450	28	16.1	9.58	43.1	46.2
	770903- 771009	85 - 115	300	36	8.3	7.00	21.0	112.9 0.40
	770806- 771009	40 - 115	750	64	11.7	9.60	72.0	159.1 0.45
	780512- 780609	40 - 80	400	28	14.3	8.30	33.2	37.0
	780609- 780905	80 - 130	500	88	5.7	9.14	45.7	278.4 0.25
	780512- 780905	40 - 130	900	116	7.8	8.22	79.4	315.4 0.25
	790904- 791109	40 - 115	750	66	11.4	7.12	53.4	89.7 0.60
	770806- 770905	40 - 70	300	28	10.7	9.60	28.8	46.2
	770905- 771009	40 - 70	0	34	0	10.90	0	112.9 0.18
	770806- 771007	40 - 70	300	62	4.8	11.32	34.0	159.1 0.21
B	780512- 780609	40 - 100	600	28	21.4	7.85	47.1	37.0 1.27
	790904- 791109	40 - 70	300	66	4.6	7.35	22.1	89.7 0.25
	770806- 770903	40 - 55	150	28	5.4	9.50	14.3	46.2
	770903- 771008	40 - 55	0	35	0	11.60	0	112.9 0.09
	770806- 771008	40 - 55	150	63	2.4	12.47	18.7	159.1 0.12
	780512- 780609	40 - 55	150	26	5.4	8.88	13.3	37.0
	780609- 780905	40 - 55	0	88	0	9.73	0	278.4 0.04
	780512- 780905	40 - 55	150	116	2.4	8.82	13.2	315.4 0.04
	790904- 791006	40 - 55	150	32	4.7	7.70	11.6	46.3
	791006- 791109	40 - 55	0	34	0	7.75	0	43.4 0.13
C	790904- 791109	40 - 55	150	66	2.3	7.73	11.6	89.7 0.13

Table 20 A table for the estimation of recharge by the ^3H -tagging method, Tärnsjö.

Stn.	Period	Depth interval, (cm)	Shift, x (mm)	Time, t. (days)	dx/dt (mm/d)	CMC, θ (vol. %)	Rech. R $\theta \cdot x$ (mm)	Prec. (mm)	r=R/P
D	770807-771009	40 - 85	450	63	7.1	10.59	47.7	112.9	0.42
	780512-780609	40 - 85	450	28	16.1	8.28	37.3	37.0	
	780609-780905	85 - 100	150	88	1.7	6.72	10.1	278.4	0.15
	780512-780905	40 - 100	600	116	5.2	8.17	49.0	315.4	0.16
	790904-791006	40 - 85	450	32	14.1	6.72	30.2	46.3	
	791006-791104	40 - 85	0	34	0	7.76	0	43.4	0.34
	790904-791104	40 - 85	450	66	6.8	7.22	32.5	89.7	0.36
	770807-770911	40 - 55	150	25	6.0	9.50	14.3	46.2	
	770911-771009	40 - 55	0	38	0	11.60	0	112.9	0.09
	770807-771009	40 - 55	150	63	2.4	12.47	18.7	159.1	0.12
	780512-780609	40 - 85	450	28	16.1	8.28	37.3	37.0	
	780609-780905	85 - 115	300	88	3.4	7.96	23.9	278.4	0.19
E	780512-780905	40 - 115	750	116	6.5	8.25	61.9	315.4	0.20
	790904-791006	40 - 55	150	32	4.7	7.70	11.6	46.3	
	791006-791109	40 - 55	0	34	0	7.75	0	43.4	0.13
	790904-791109	40 - 55	150	66	2.3	7.73	11.6	89.7	0.13

Table 21 Groundwater recharge estimated by the water balance method, Tärnsjö, station C.

Stn	Period	Number of days	ET (mm)	dS (mm)	P (mm)	R (mm)	r
C	770806-0903	29	69.90	-16.80	46.20	1.10	
	770903-1008	<u>35</u>	61.04	32.70	112.90	<u>19.30</u>	
		64				20.40	0.18
	780512-0609	28	91.10	-56.10	37.00	2.00	
	780609-0905	<u>88</u>	221.10	42.50	278.40	<u>13.80</u>	
		116				15.80	0.05
	790904-1006	32	34.97	-18.80	46.30	15.60	0.34

Tables 22 and 23 show the recharge estimates for Jädraås area using the ^3H -tagging method and the water balance method respectively. The 1976 estimates using the tritium-tagging method are 56.9 mm and 66.8 mm for the period 760612 - 761112, which shows quite good agreement for the two closely situated plots with similar soil material. The 1977 results of 27.8 mm and 22.6 mm show even closer values. The results of the water balance method are somewhat higher than those by the tritium-tagging method.

Table 22. A table for the estimation of recharge by the ^3H -tagging method for Jädraås area.

Stn. Period	Depth interval, (cm)	Shift, x (mm)	Time, t. (days)	dx/dt (mm/d)	CHC, θ (vol. %)	Rech. R $\theta \cdot x$ (mm)	Prec. (mm)	$r=R/P$
IHII	760612-760922	40 - 85	450	102	4.4	6.55	29.5	94.0
	760922-760929	85 - 145	600	7	85.7	7.21	43.3	2.0
	760929-761009	145 - 85	-600	10	60	7.00	-42.0	12.0
	761009-761112	85 - 115	300	34	8.8	9.37	28.1	127.0 0.18
	760612-761112	40 - 115	750	153	4.9	7.59	56.9	235.0 0.24
	770604-770621	40 - 55	150	17	8.8	5.21	7.8	19.0
	770621-770708	55 - 85	300	15	20.0	7.45	22.4	77.0 0.32
	770604-770708	40 - 85	450	32	14.1	6.18	27.8	96.0 0.29
	760612-761007	40 - 130	900	117	7.7	9.01	81.1	108.0
	761007-761113	130 - 115	-150	37	4.1	8.21	-12.3	127.0 0.21
IHV	760612-761113	40 - 115	750	154	4.9	8.91	66.8	255.0 0.28
	770604-770708	40 - 70	300	32	9.4	7.53	22.6	96.0 0.24

The groundwater recharge values obtained by the tritium-tagging method and by the water balance method for the three areas are summarised in table 24. Some general comments can be deduced from the table.

Table 23. Recharge estimates by the water balance method for Jädraås area. (Obtained from Per Erik Jansson, 1977)

Date	Precipitation accum. (in mm)	Evapotranspir- ation, accum. (in mm)	Recharge (in mm)	
			Accum.	For obs. period
760922	94	16	31	
760929	96	20	38	
761007	108	25	42	
761009	108	27	42	
761112	235	40	127	
761113	235	40	139	139
770604	636	120	560	
770615	655	150	568	
770618	655	158	569	
770627	691	174	572	
770707	732	195	590	30

Referring to table 24, and Skediga area, the agreement of the recharge estimates for plots A and B is quite satisfactory. In the layered soil, plot C, the ^3H -tagging method gives high values of recharge as compared to that of the water balance method. For a very coarse material, plot D, the reverse is true i.e. the water balance method gives a higher recharge estimates. We can also see that the recharge estimates for periods that are too short leads to erroneous results. (the 1978 results for plot C).

The recharge estimates for Tärnsjö area by the two methods are available only for plot C. A comparison of the results by the two methods shows that the variation between the results of the two methods is 6% for 1977, 1% for 1978 and 11% for 1979. Considering the areal variation of the soil properties, the agreement of the two methods is very satisfactory.

In Jädraås area, the comparison of the two methods shows that the water balance method for 1976 gives extremely high values for one of the stations. Such errors are expected when using the water balance method which contains evapotranspiration term and this factor is difficult to estimate accurately. Further more the recharge estimates are point values calculated using areally estimated evapotranspiration values. This can also be one source of error. The 1977 results on the other hand show good agreement both for the different methods and stations, an average of 25.2 mm for the tritium-tagging method 30.0 mm for the water balance method.

Table 24. Recharge estimates by the ^3H -tagging and the water balance method for the three sites.

Site	Plot	Period	Days	^3H -tagging	H_2O -balance	(^3H) r	(wb) r
Skediga	A	760505-1009	126	95.5	88.3	0.58	0.54
		760606-1010	126	91.2	96.4	0.57	0.60
	C	760606-1023	139	125.2	89.7	0.77*	0.55
		770630-0829	60	129.6	110.8	0.74	0.63
		780509-0606	28	56.8	26.3	2.00*	0.93
		790903-1204	92	138.8	83.6	0.85	0.51
	D	760606-1024	141	66.6	86.3	0.44	0.53
		770630-0730	31	32.6	97.7	0.21	0.64
		780505-0828	115	36.7	173.9	0.13	0.61
		790903-1204	92	29.2	85.5	0.18	0.53
	Tärnsjö	770806-1008	64	18.7	20.4	0.12	0.18
		780512-0905	116	13.2	15.8	0.04	0.05
		790904-1006	32	11.6	15.6	0.25	0.34
Jädraås	IHII	760612-1112	171	56.9	139.0	0.18	0.24
		770604-0708	34	27.8	30.0	0.29	0.30
	IHV	760612-1112	171	66.8	139.0	0.20	0.54
		770604-0708	34	22.6	30.0	0.24	0.31

r = R/P (R = recharge, P = precipitation)

9. DISCUSSION AND CONCLUSION

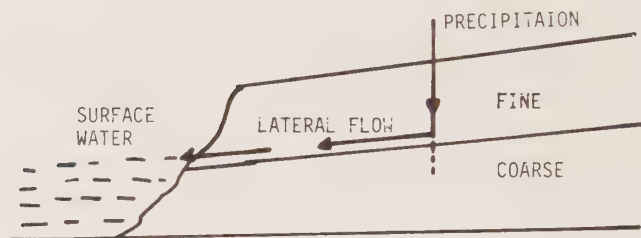
Both the ^3H -tagging method and the water balance method give reliable results when the soil profile is homogeneous. Stratified soil profiles such as coarse-over-fine or fine-over-coarse layered soils have a pronounced effect on both the moisture movement and tracer displacement.

In a fine-over-coarse layered soils, the tracer is delayed in the upper part of the boundary because of the higher tension compared to that in the lower boundary in the coarser material. This delayed vertical movement as observed in Skediga, gives lower recharge values. According to L. J. Andersens (1974) investigation of soil moisture movement in glaciofluvial outwash material, the velocity of the soil moisture front is 3 - 3.5 m monthly and the actual water velocity is 4.5 m yearly. Although the difference in velocity between the front and the water is not quite understood, the values given are higher than those encountered in the heterogeneous material of Skediga area where the highest observed short time velocity is 26 mm per day which is equal to only 0.78 m per month.

If layering occurs in inclined soil profiles as shown in figure 53, there can occur a fast lateral flow or interflow which in turn can contribute to direct surface runoff without reaching the groundwater storage. This fast lateral flow is also observed in a continuing investigation which shall be reported in a forthcoming paper.

It is also observed that, due to the high areal variabilities of recharge measurements by point observations, it is essential that as many observation points as possible must be used in the study of recharge for a catchment area. An estimate based on very few observation points can give erroneous results.

Fig. 53. Lateral flow in a stratified soil layer



Other workers using tritium-tagging method for recharge measurements have also observed a pronounced areal variability. Bahadur et al (1977) studied groundwater recharge through a soil material of 0.5 - 20% coarse sand, 37 - 53% fine sand and of varying amount of silt and clay contents. The recharge values they obtained varied from about 12.5% to 45.0% in these sandy loam soils. Groundwater recharge values estimated by the ^3H -tagging method for the Lower Maner Basin in India, by Athavale et al (1978) showed values ranging from 5 cm to 24 cm at the different observation points. In the present investigation the recharge values varied between 13 mm per day and 79 mm per day for Tärnsjö area where all the different soil types are represented. In Jädraås where the soil types are relatively similar, the variations in recharge values are not so great (56.9 mm and 66.8 mm for 1976, 27.8 mm and 22.6 mm for 1977) In Skediga where the soil is stratified (plot C) the variations are still higher.

The effect of layering on tracer distribution is also observed. When the soil layer is stratified, the tracer distribution becomes distorted and does not follow the usual Gaussian distribution. The tracer profile becomes narrower when it passes from a bigger to a smaller layer of finer grain size and broadens when the layering is reversed. (Blume et al, 1967)

The soil moisture content summations for the different layers and times of observations as shown in figs. 38, 39, 43, 44 and 45, clearly indicate two distinct periods of "filling" (from about August/September to about April/March) and of "emptying" (from about April/March to about August/September). This process of filling and emptying is very fast just after snowmelt. This fact combined with the initial fast movement of tritium-tagged water during the same periods leads to the conclusion that most of the yearly recharge is made up of the short time recharge after the snow melt. This is further elucidated by using the short term recharge to estimate the total evapotranspiration loss. 1976 recharge values for Skediga and for plots A, B and D are 95.5 mm, 91.2 mm and 21.2 mm with an average value of 69.3 mm. Similarly the 1978 recharge values for Tärnsjö and for plots A to E are 79.4 mm, 47.1 mm, 13.2 mm, 49.6 mm and 61.9 mm with an average value of 50.2 mm. These recharge values are for the period May/June to about September. The differences between these short time recharge values and the corresponding total precipitation values (491 mm for Skediga and 502 mm for Tärnsjö) are 347 mm and 452 mm. These values do not fall far outside the range for the total evapotranspiration loss for these areas as calculated by B. Eriksson (1980) and which lie between 400 mm and 500 mm.

It is observed, especially in Skediga area, that in a stratification of fine-over-coarse sequence, it seems that the ^3H -tagging method gives a higher recharge values than the water balance method. In a very coarse material the water balance method gave higher recharge estimates compared to the tritium method.

Observation periods of a very short duration also gave values which are unacceptable. The 1978 results for the 28 days between the 9th. of May and the 6th. of June are good examples.

In general one can conclude that the tritium method is inaccurate when:
 " (a) Large inhomogenities of the soil substrate occur perpendicular to the flux, causing seepage pathways with significantly different velocities.

(b) The soil water suction gradients at the injection depth show a large spacial variability when the injection is performed. In such cases the injection liquid moves differently already at the starting point.

(c) The soils are stony or sandy, so that problems arise for sampling cores in the subsoil, and a continuous activity profile might not be gained.

Avoiding these problems and taking account of the restrictions mentioned, the HT0-tracing method may be taken as a well suited method" (Kreutzer et al, 1980).

One of the drawbacks of the water balance mehtod is the difficulty of estimating the correct value of evapotranspiration loss since there lacks a direct method for measuring this variable.

As far as the soil moisture content determinations are concerned, the neutron scattering method using a probe with a fast neutron source and thermal neutron detector is the best method for insitu monitoring of soil moisture variations with time. For dry conditions with homogeneous sandy formations, the air suction method with molecular sieve is best suited. The new TDR (Time Domain Reflectometer) method makes it possible to follow the movement of unfrozen moisture in freezing soil during the winter season. The soil moisture profiles are essential not only to see the vertical moisture movement but also to observe the stratification or material distribution of the soil column in question.

10 SUMMARY

In this investigation an attempt has been made to study soil moisture movement in the unsaturated zone and estimate groundwater recharge to the unconfined esker aquifer using the tritium-tagging method. The recharge estimates by the tritium method are compared to those estimated by the water balance method. The tritium-tagging method is based on the principle that soil moisture moves in layers without mixing with the moisture ahead of it, the principle known as "piston flow". This makes it possible to make a water balance for soil column between the injection point and the peak position.

The soil moisture content variations are monitored using the neutron moderation method and the evapotranspiration values are estimated using the Penman formula which are then reduced to the actual values by a soil moisture factor.

The percolation rates vary both in space and time. The early percolation rates just after snow melt can be as high as 21 mm/day (in Tärnsjö) and 26 mm/day (Skediga). The late summer percolation rates can be as low as 2.1 to 2.3 mm/day. But the average values for the areas agree quite well (8.7 mm/day for Skediga and 8.6 mm/day for Tärnsjö). The groundwater recharge values show big areal variations due to the inhomogeneity of the soil but the average values for the areas do not differ much. For example the 1976 recharges for Skediga and Tärnsjö are 61.9 mm and 50.2 mm with the R values of 17% and 16%. Comparison of the methods shows that in stratified soils with fine-over-coarse sequence the tritium method gives higher recharge values and in very coarse materials the reverse is true. It is also observed that very short observation periods give unreliable recharge values.

As far as the sampling methods are concerned, the porous cup method does not function very well in esker materials due to difficulties in installation and the poor contact between the cells and the soil material. In such materials and relatively dry conditions the soil air

suction method with molecular sieve-bed is a good method. Even soil coring is difficult in coarse materials.

In general combined soil moisture studies and tracer applications give much information about the soil conditions and recharge processes.

11.1 Tracer elements in hydrology

Water movement in the soil, especially in the unsaturated zone is a very complicated process. It is the opinion of many workers that the problems of geohydrology, for a long time have been difficult to solve because the methods used have been heavy, expensive, and time consuming. As a result the use of tracer elements has become more and more popular. Already as early as 1869, tracers such as dyes and salts have been used to determine hydraulic connections. There are many kinds of tracers some of which are shown in the chart below:

Tracer	Example	Area of application
1. Commercial tracers	a) <u>Dyes</u> : Uramine, Rhodamine B b) <u>Salts</u> : NaCl, KCl, NH ₄ Cl, NaI (strong electrolytes) LiCl ₂ , NH ₄ Br, NH ₄ , SO ₄ c) <u>Radioactive substances</u> : Tritium (³ H or T) d) <u>Stable metal chelates</u> : ⁵¹ Cr-EDTA	Water movement and direction of flow
2. Naturally occurring non-radioactive substances	Deuterium (D), ¹⁸ O, ³⁴ S, ³² S, Cl ⁻	Origin of water
3. Naturally occurring radioactive substances	Carbon -14 (¹⁴ C), Tritium (³ H or T)	Dating

The main advantages of tracers to solve geohydrologic problems is that:

- 1) They do not change the natural conditions
- 2) They are easy to apply
- 3) They have relatively low cost

1.2 The selection of a tracer

There are several criteria in the choice of a tracer.

- a) The purpose of the investigation
- b) Detectability of the tracer in low concentrations in lab and in field.
- c) The interaction of the tracer with the media, ie if the tracer is absorbed or takes part in an ion exchange process in the ground.
- d) The type of aquifer or soil to be investigated.

When considering radio-active isotopes, which are being increasingly used in hydrological studies, their half-life is of prime importance. Depending on the length of the investigation time, which can be short, medium or long term (16), an appropriate type of tracer with a similar half-life must be selected.

Some common examples are given below.

Duration	Tracer	Half-life	MPC ($\mu\text{Ci}/\text{cm}^3$)	Comments
Short term	^{131}I	8.06 days	2×10^{-6}	Too short half-life, low MPC. Otherwise widely used in oil fields, pumping tests, groundwater flow in gravelly and sandy soils.
	^{82}Br	36.0 hrs.	3×10^{-4}	Good for high flow rates. In all other respects the same like ^{131}I .
Med. term	$^{51}\text{Cr}.\text{EDTA}$	27.8 days	2×10^{-3}	No β -radiation, has high MPC, has been widely used in alluvial soils, glacio-fluvial soils, sands, gravels and karst formations.
Long term	^{36}Cl	3.08×10^5 years	6×10^{-5}	Very difficult and costly to prepare.

Some of the most widely used tracers are the environmental radio isotope tritium and/or $^{51}\text{Cr-EDTA}$, the stable environmental isotopes "Deuterium" "D" and Oxygen 18.

11.3 Tritium

Definition: Tritium (T or ^3H) is one of the isotopes of hydrogen. The ordinary hydrogen nuclide contains one proton (^1_1H); Heavy hydrogen (D) or deuterium nuclide contains one proton and one neutron (^2_1H) and the "Super heavy hydrogen", tritium (T) has one proton and two neutrons (^3_1H).

Tritium disintegrates with the emission of β -radiation of very low energy (17.6 Kev). It has a half life of 12.5 years. The MPC is $3 \times 10^{-3} \mu\text{ci}/\text{cm}^3$.

Units

Tritium units (TU): A tritium unit is defined as one atom of ^3H per 10^{18} atoms of hydrogen atoms

also $1 \text{ TU} = 7.2 \text{ dpm per litre of water}$

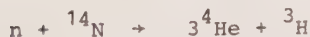
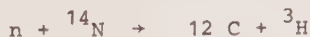
or $1 \text{ TU} = 3.2 \times 10^{-3} \mu\text{ci}/\text{cm}^3 = 3.2 \text{ ci/litre}$

A curi (Ci) = $3.7 \times 10^{10} \text{ dis/sec}$

Source_of tritium

11.3.1 Natural

The impact of high energy cosmic ray neutrons with atmospheric nitrogen producer tritium according to the reaction:



at a constant rate of $0.2 - 2 \text{ atoms}/\text{cm}^2/\text{sec}$. Cosmic ray produced ^3H atom has a very short life in the free state because it is oxidized to tritiated water HTO which falls down with rain.

The pre-1954 cosmic ray produced tritium distribution in rain water is estimated to be 5-10 TU throughout most parts of the world varying with latitude and with distance from the coast.

11.3.2 Tritium produced since 1952 (Man-made tritium since 1952)

Starting in 1952, due to the series of thermonuclear explosions high quantities of ^3H have been injected into the stratosphere which eventually reached the troposphere from where it fell down with rain. The peaks in tritium fallout increased after successive nuclear detonations i.e. 1954, 1956, 1958/9, 1962/83. In 1963 the ^3H concentration in rain water rose as high as 10 000 TU in the extreme north. In the southern hemisphere the increase of ^3H in the atmosphere is very little, the max being about 90 TU (Pretoria, 1963).
Note! Latest records of tritium in rain water show

- a) Seasonal variation with a maximum in the late spring and summer i.e. May-June and a minimum during the winter months Nov-Dec
- b) Latitudinal variation with an increase towards the Poles
- c) In general the tritium in rain water is much lower in the southern hemisphere

Some definitions

Tritium profile = The distribution of tritium along a soil column is called tritium profile

TU meter = Tritium concentration x weight/unit mass of water x the thickness of the profile

If there is a long term data on the precipitation and its tritium content, the natural tritium profile at any particular place can give a good estimate of the rate of the downward flow of water (infiltration rate) and also one can say if the infiltrated water is old (pre 1952) or recent (after 1952).

11.3.3 Reasons for the choice of tritium

- 1 Tritiated water (HTO) is identical with water
- 2 Detection sensitivity of tritium is high
- 3 Very low risk for radiotoxicity (low emission energy of radiatum. MPC = $3 \times 10^{-3} \mu\text{Ci}/\text{cm}^3$)
- 4 Tritium is insignificantly retarded by water bearing formations of sandy gravelly type
- 5 The half life (12.3 years) is long enough for ground water investigation
- 6 Its half life is long enough to store it in the lab
- 7 Tritium has a low cost

The main disadvantage of using tritium is the difficulty to perform any field detections.

11.4 SOME METHODS FOR THE ESTIMATION OF EVAPOTRANSPIRATION IN NATURE

(a) THE TANK OR THE PAN METHOD

Evaporation tanks or pans are usually circular and are made in various sizes and of different metals; they can be floating types or types that are sunk into the ground. A class A evaporation pan 1.22 meters in diameter and 0.254 meters in depth is used by the U.S. Weather Bureau (Rodda et al, 1976)

The Colorado sunken pan is 3 feet square and 18 inch deep used by the U.S Weather Bureau and the GGI-3000 pan used in USSR and with an area of 0.3 m^3 and 60 cm depth are two other examples. Floating pans are meant to simulate reservoir evaporation by floating a pan in water. The evaporation loss from such pans is equal to the amount of water needed to bring the level of water in the pan to that in the surrounding water. Due to relatively small size of pans a condition is created which allows greater amounts of advected heat from the atmosphere to be absorbed by the water in the pan through the sides and bottom of the pan (Wilson, E.M. 1969). Thus the pan method overestimates evaporation and a pan coefficient must be applied. The value of the coefficient depends upon the size of the pan and can range from 0.65 to 1.12.

(b) LYSIMETERS

Lysimeters are usually made by filling an oil drum with the type of soil in question undisturbed if possible and then installing it so that its rim is level with the ground. Water draining through the lysimeter is conducted into measuring cylinder placed further away in a pit. The difference between the infiltrated water plus precipitation (if any) and the drained water gives the evapotranspiration loss. Hence,

$$E = (P + IR) - ID \quad (31)$$

where, E = evaporation; P = precipitation; IR = infiltrated water
ID = drainage water.

(c) ATMOMETER

An atmometer consists of a porous material with a reservoir of distilled water underneath. The porous material simulates the evaporating surface and is used instead of a pan (Rodda et al 1976).

(d) THE WATER BUDGET METHOD

The water budget or the storage equation can be written as,

$$ET = P + SI + UO - SO \pm SR \quad (32)$$

where, ET = evapotranspiration

P = precipitation

SI = surface inflow

UO = underground outflow

SO = surface outflow

SR = change in surface or subsurface storage.

If the catchment area is defined all the terms on the right hand side of the expression can be estimated. The term SR is negligible for very long observation periods. This method gives a good approximation of the evapotranspiration loss in nature.

(e) THE ENERGY BUDGET METHOD

The primary source of energy at the surface of the earth is solar energy. The net radiation flux density absorbed by the earth's surface can be expressed as,

$$R_n = (1 - \alpha) (R_s + R_d) + \epsilon(R_L - \sigma T^4) \quad (33)$$

where, α = the surface albedo

$(R_s + R_d)$ = direct and diffused short wave solar radiation
in the wavelength range from 0.3 to 3 microns

ϵ = the thermal emission coefficient of the surface

R_L = long wave radiation

σ = thermal absorbtivity of the surface

T = the absolute temperature of the surface in $^{\circ}\text{K}$.

This net radiation absorbed by the surface is totally dissipated in the following three ways.

(a) to raise the temperature of the soil and of the plants as ground heat flux density, G

(b) to heat the air above the surface of the earth as sensitive heat flux density, H and

(c) for evaporation as latent heat flux density, E

Thus the energy balance method can be written as,

$$R_n = G + H + E \quad (34)$$

The energy budget method involves a great deal of instrumentations and cannot readily be used without much data (Wilson, E.M., 1969)

(f) THE MASS TRANSFER METHOD

The mass transfer approach is based on the assumption that there exists a water vapour pressure gradient in the air layer above the surface of the earth and considers evaporation in analogy to diffusion process. The equation is written as,

$$E = (LE \cdot \rho \cdot \omega / p) K_E \cdot de/dz \quad (35)$$

where, LE = latent heat of vaporization

ρ = density of air

ω = relative molecular weight of water with respect to air

p = barometric pressure

K_E = eddy diffusivity of water vapour

de/dz = vertical water vapour pressure gradient

Since K_E is not constant but dependent on wind conditions the application of the mass transfer formula is of no practical importance.

(g) THE AERODYNAMIC APPROACH

Evaporation can only continue if the vapour produced by the available energy is also removed. This requires the existence of moisture vapour gradient and wind. There are several approaches in this line. One such example is shown below.

$$E = \alpha(1 + \beta u)(e_s - e_d) \quad (\text{Rodda et al 1976}) \quad (36)$$

where, α and β are constants; u is the wind speed at height h and e_s and e_d vapour pressures at two heights.

(h) THE THORNTHWAITES FORMULA

The Thornthwaite formula is built only on mean temperatures and is good for stations that lack a complete meteorological data. This method was devised to estimate the potential evapotranspiration from short and close set vegetation with an adequate water supply in the latitudes of the United States (Wilson, 1969). The basic equation is of the form,

$$PET = 1.6(10T/J)^a \quad (37)$$

where, $J = \sum_{1}^{12} j$ (sum for the 12 months)

$$j = (t_n/5)^{1.514} = \text{monthly heat index}$$

$$t_n = \text{Average monthly temperature in } ^\circ\text{C}$$

$$a = 0.016J + 0.5 \text{ (in its simplified form)}$$

This method is empirical, complicated and needs a monogram for its solution but for project studies the method is a good complement to the Penman method.

(i) THE CROWE AND (j) THE TURC FORMULAS

The two other methods mentioned in the literature are the Crowe and the Turc formulas. The former uses the daily maximum temperatures in $^\circ\text{F}$ and the latter precipitation and mean air temperature. The Turc estimates evaporation on a yearly basis.

11.5 Tables 4, 5, 6, 7A - D, 8, 9A - D, 10A - D, 11 and 12

Table 4. Soil moisture contents by volume % for Jädraås profiles.

TIME	10	20	30	40	55	70	85	100	115	130	145	160	175	190	205
751118 12100	20.70	15.458	9.645	8.199	7.206	6.842	6.951	7.389	8.155	8.229	8.517	7.475	8.991	6.916	6.515
751222 12100	24.87	9.842	8.737	7.403	6.667	6.171	6.042	6.441	7.297	7.445	7.681	7.043	8.787	7.018	7.406
760122 12100	22.63	10.842	10.852	8.323	6.451	5.425	5.319	5.612	6.337	6.345	6.526	6.275	7.407	6.094	6.785
760223 12100	45.01	20.416	10.675	8.361	6.791	5.915	5.496	6.317	6.496	5.661	5.806	5.823	6.755	6.263	6.269
760323 12100	53.61	25.242	13.733	12.395	9.341	7.175	6.082	6.483	5.894	5.206	5.572	5.744	5.766	5.209	6.114
760426 12100	55.34	25.176	13.675	12.402	9.419	7.127	6.072	6.727	6.108	5.126	5.417	4.999	5.926	5.053	5.853
760618 12100	34.51	25.142	16.361	15.899	14.791	12.563	10.496	9.531	9.601	7.706	7.122	6.429	5.748	5.714	5.389
760427 12100	25.28	20.858	12.784	11.340	10.505	9.218	7.635	7.160	7.196	6.867	7.123	6.429	5.848	6.466	6.493
760504 12100	26.19	21.311	13.389	11.784	10.596	9.011	7.335	6.674	6.895	7.370	7.227	6.510	6.819	6.473	6.663
760515 12100	19.40	18.226	11.516	10.056	9.854	9.045	8.557	8.045	7.972	7.370	7.227	6.510	6.819	6.473	6.663
760524 12100	16.20	14.853	9.401	7.994	7.126	7.653	6.833	7.097	6.478	9.147	7.370	7.227	6.819	6.473	6.663
760524 12100	16.20	14.853	9.401	7.994	7.126	7.653	6.833	7.097	6.478	9.147	7.370	7.227	6.819	6.473	6.663
760601 12100	14.26	13.290	8.743	7.159	6.444	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760608 12100	13.18	11.911	7.876	6.626	6.208	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760616 12100	18.84	11.182	10.447	8.420	6.473	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760621 12100	16.76	15.025	9.081	8.420	6.473	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760629 12100	14.18	12.892	8.231	6.890	6.214	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760712 12100	11.87	11.134	7.316	5.389	5.138	4.958	5.031	5.418	5.720	5.720	5.937	5.223	5.920	5.002	6.381
760721 12100	7.21	7.347	3.725	3.443	3.138	4.958	5.031	5.418	5.720	5.720	5.937	5.223	5.920	5.002	6.381
760804 12100	16.45	11.988	8.831	8.420	6.473	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760827 12100	6.98	11.681	8.443	8.420	6.473	6.088	5.011	5.144	5.720	6.193	6.844	7.145	9.014	7.81	9.166
760901 12100	10.53	14.877	9.945	8.126	7.666	6.816	6.253	5.262	4.637	4.747	4.711	4.313	4.681	4.276	5.343
760921 12100	18.46	12.243	10.996	9.234	8.279	7.911	8.021	8.940	10.631	11.991	12.505	9.969	11.697	10.667	11.657
760922 12100	12.29	14.313	9.158	7.437	6.996	6.615	6.450	6.229	6.815	6.888	7.291	6.742	8.280	6.266	7.108
761013 12100	21.59	18.245	11.955	10.201	9.237	8.566	7.971	8.044	7.552	7.972	8.280	6.266	7.423	5.814	5.595
761109 12100	22.76	19.245	12.351	10.402	9.673	8.981	8.491	10.602	12.169	13.007	12.024	10.456	12.075	12.470	6.252
761123 12100	19.72	15.181	9.726	7.714	6.996	6.615	6.450	6.229	6.815	7.972	7.972	7.253	9.927	7.146	7.963
761221 12100	26.59	18.659	10.043	8.156	7.559	6.868	6.705	7.104	7.649	7.794	7.890	7.249	8.763	7.113	6.963
770121 12100	30.32	18.999	7.737	6.361	5.745	5.456	5.392	5.792	6.325	6.763	5.927	5.032	6.584	5.509	6.099
770223 12100	33.84	18.870	8.715	6.297	5.468	5.456	5.392	5.792	6.325	6.763	5.927	5.032	6.584	5.509	6.099
770319 12100	33.74	18.084	9.207	6.144	5.063	4.991	4.847	5.179	5.612	5.540	5.320	5.304	6.541	5.208	6.000
770319 12100	44.76	22.987	11.408	7.925	6.424	5.457	5.392	5.792	6.325	6.763	5.927	5.032	6.584	5.509	6.099
770404 12100	50.53	26.327	11.573	7.792	6.740	6.424	5.457	5.392	5.792	6.325	6.763	5.927	6.541	5.208	6.000
770428 12100	57.36	31.169	15.227	11.402	10.044	8.573	8.345	8.646	9.277	10.241	10.905	10.905	13.291	10.648	8.199
770521 12100	30.90	24.390	13.733	10.175	9.200	8.910	8.941	10.199	11.707	12.741	13.210	10.905	13.291	10.648	8.199
770527 12100	15.36	15.165	9.593	7.573	6.844	6.407	6.199	6.734	6.990	6.951	7.245	6.595	6.155	6.844	7.464
770603 12100	12.97	13.218	9.721	7.245	6.470	5.991	5.917	6.027	6.618	6.581	6.618	6.175	6.927	6.063	6.927
770609 12100	12.97	12.155	10.446	6.674	7.871	7.082	7.229	6.954	6.516	6.516	6.516	6.063	6.927	6.063	6.927
770622 12100	17.736	13.901	8.598	7.211	6.446	6.094	6.131	5.871	5.908	5.963	5.945	5.677	6.477	5.677	6.477
770707 12100	19.776	12.725	11.434	9.502	8.758	8.164	7.978	7.457	6.082	6.974	6.082	5.690	5.677	5.677	6.477
770713 12100	12.820	15.823	9.844	8.293	7.406	7.293	7.073	7.774	8.957	9.622	9.438	8.070	8.718	7.718	8.718
770713 12100	12.820	12.618	8.451	7.266	6.529	6.271	6.161	6.603	7.302	7.744	7.671	7.118	8.718	7.718	8.718
770726 12100	18.140	18.557	11.930	9.370	8.196	7.929	8.206	9.481	11.317	12.859	12.969	10.657	11.317	10.657	11.317
770825 12100	20.128	16.990	10.625	8.899	7.704	7.011	6.628	6.791	7.120	7.120	7.339	6.155	5.823	5.823	6.155
770819 12100	13.193	12.675	9.546	6.964	6.191	6.007	5.824	5.897	6.192	6.192	6.192	5.654	6.192	5.654	6.192
770825 12100	11.367	10.921	7.687	6.470	5.890	5.495	5.495	5.495	5.890	5.890	5.890	5.495	5.890	5.495	5.890

Table 5. Soil moisture contents by vol.% for the different depths and periods for Jädraås plot IH VA 1.

PERIOD	10	20	30	40	50	60	70	80	90	100	110	120	140	160	180
760612- 761007	14.08	13.00	8.70	7.67	6.64	6.20	6.05	5.99	6.15	6.25	6.45	5.75	6.84	5.66	6.62
761007- 761113	22.18	18.76	12.15	10.40	9.58	9.07	8.83	9.32	9.85	9.78	9.67	8.21	9.81	9.19	7.59
770604- 770618	16.39	15.19	9.58	7.94	7.15	6.74	6.57	6.48	6.57	6.50	6.57	6.13	7.60	6.34	7.04
770618- 770627	17.73	13.90	8.59	7.21	6.47	6.09	8.16	10.70	5.91	5.98	5.95	5.50	6.54	5.72	7.25
770627- 770607	18.65	16.77	10.64	8.90	8.08	9.43	7.53	7.62	7.97	7.85	7.65	6.72	10.21	8.55	7.18
770604- 770627	16.83	14.76	9.25	7.70	6.92	6.52	6.43	6.28	6.35	6.33	6.36	5.92	7.07	6.03	7.14

Table 6. Soil moisture contents by vol.% for the different depths and periods for Jädraås plot IH VA 2.

PERIOD	10	20	30	40	50	60	70	80	90	100	110	120	140	160	180
760612- 760922	14.00	13.01	8.73	7.78	6.69	6.25	6.08	5.96	6.05	6.16	6.34	5.60	6.59	5.51	6.47
760922- 760929	15.87	13.99	9.16	7.44	6.74	6.19	6.05	6.23	6.82	6.89	7.29	6.74	8.28	6.27	7.11
760929- 761009	21.59	18.27	11.95	10.20	9.29	8.56	7.97	8.04	7.53	6.55	6.51	5.96	7.42	5.81	6.66
761009- 761112	22.18	18.76	12.15	10.40	9.58	9.07	8.83	9.32	9.85	9.78	9.67	8.21	9.81	9.19	7.59
770604- 770621	16.83	14.76	9.25	7.70	6.92	6.52	6.43	6.28	6.35	6.33	6.36	5.92	7.07	6.03	7.15
770621- 770708	18.34	15.82	9.96	8.34	7.54	8.32	7.06	7.03	7.28	7.23	7.06	6.32	8.38	7.14	7.22

Tables 7A - 7D Soil moisture contents by vol. % for Skediga plots

7A

TIME	10	20	30	40	55	70	85	100	115	130	145	160	175
76007 12:00	52.91	32.870	19.048	17.706	15.023	11.687	9.143	7.595	6.391	6.150	6.254	6.254	6.701
760615 12:00	4.95	19.667	18.348	13.221	11.949	11.004	9.145	8.198	7.471	7.444	7.652	7.971	8.075
760704 12:00	19.71	16.073	10.287	9.825	9.049	8.617	7.677	7.016	7.225	7.399	7.503	7.956	8.108
760824 12:00	13.67	11.660	7.915	7.636	7.391	7.322	7.276	6.623	6.727	6.937	7.042	7.252	7.336
760907 12:00	19.92	12.326	8.101	8.101	7.871	7.786	7.726	6.666	6.806	6.946	7.136	7.136	7.875
761004 12:00	34.41	25.619	13.946	13.034	12.543	11.689	8.893	8.016	7.875	7.910	7.524	7.524	7.937
761104 12:00	42.41	25.516	13.377	14.568	12.373	10.982	8.802	8.016	7.875	7.713	7.220	7.220	7.937
761204 12:00	50.92	28.950	17.962	18.208	15.850	14.372	11.743	10.677	10.300	10.147	9.701	9.761	9.797
770208 12:00	39.39	28.423	14.280	14.716	12.446	11.372	9.847	7.848	7.848	7.848	7.848	8.000	8.246
770321 12:00	42.14	34.310	17.101	15.209	13.546	12.546	11.668	9.987	9.987	9.987	9.987	9.987	9.987
770418 12:00	33.49	18.638	17.296	16.874	16.874	14.616	12.288	10.806	10.806	10.806	10.806	10.806	10.806
770523 12:00	33.29	23.652	14.277	13.849	11.907	11.025	8.624	8.059	8.412	8.534	8.730	8.518	8.891

7B

TIME	10	20	30	40	55	70	85	100	115	130	145	160	175	190	205
76007 12:00	48.644	36.251	27.175	24.935	19.492	11.775	9.950	9.168	7.526	7.191	7.435	9.903	11.913	12.919	6.991
760615 12:00	15.101	22.387	19.376	19.410	19.479	14.729	10.323	10.177	7.514	6.894	6.994	8.901	11.447	12.475	9.700
760704 12:00	31.397	28.287	22.747	24.582	24.075	18.311	14.729	14.729	14.305	13.957	14.061	16.151	18.521	21.310	14.400
760824 12:00	15.301	15.035	12.388	11.627	11.627	10.626	9.716	8.942	6.172	6.022	6.102	7.296	8.096	8.531	6.242
760907 12:00	15.867	15.441	12.389	11.397	11.397	10.536	7.055	5.929	5.929	5.900	5.845	6.775	7.686	7.686	5.865
761004 12:00	19.922	17.673	14.658	13.797	16.270	10.161	7.944	7.563	6.299	5.772	5.948	6.755	7.668	7.668	5.772
761104 12:00	19.728	20.385	22.444	24.022	24.375	13.074	7.997	7.575	5.992	5.711	6.027	7.871	7.610	8.273	5.781
761204 12:00	27.706	27.008	24.584	29.132	25.974	22.363	22.751	14.889	9.254	6.957	6.252	7.380	8.825	9.143	5.751
770208 12:00	31.995	43.808	34.183	27.850	24.762	21.817	22.218	18.746	13.333	9.961	9.643	14.220	13.630	12.008	6.574
770321 12:00	30.638	31.253	24.595	20.310	20.846	18.668	18.479	16.679	13.465	10.781	10.604	15.686	16.357	17.134	9.875
770418 12:00	30.391	21.599	20.210	19.977	20.846	20.453	20.257	15.517	12.056	9.865	9.689	13.432	15.160	17.467	11.967
770523 12:00	32.391	32.564	25.337	25.091	25.867	23.642	20.357	16.087	13.065	10.009	10.045	14.269	14.917	17.248	10.713
770620 12:00	35.334	30.764	24.567	25.991	25.556	20.185	21.281	16.087	12.065	10.009	10.045	14.269	14.917	17.248	10.713
770708 12:00	32.648	26.141	22.592	23.452	23.794	19.517	19.447	16.768	12.077	10.279	10.188	14.267	15.524	17.644	10.713
770920 12:00	27.690	26.114	20.298	20.581	20.793	17.612	17.046	14.678	12.077	9.871	9.388	12.547	15.362	17.834	10.713
771004 12:00	32.510	27.734	22.894	22.916	22.245	16.907	16.094	14.756	11.478	9.660	9.625	12.547	15.362	17.834	10.713
771121 12:00	32.931	27.734	22.894	22.916	22.245	16.907	16.094	14.756	11.478	9.660	9.625	12.547	15.362	17.834	10.713
771204 12:00	33.199	30.764	23.659	24.193	25.112	19.279	17.405	15.254	14.087	11.942	11.112	16.279	15.992	18.031	10.981
780505 12:00	40.617	39.005	33.659	31.585	26.352	21.543	21.804	17.972	12.967	11.041	9.804	12.704	17.454	17.972	12.779
780623 12:00	36.730	26.730	22.181	21.073	22.744	20.099	17.053	15.427	11.973	9.099	9.840	14.013	14.720	17.377	11.113
780823 12:00	34.778	29.493	22.181	21.073	22.744	20.099	17.053	15.427	11.973	9.099	9.840	14.013	14.720	17.377	11.113
781011 12:00	32.246	29.227	22.430	23.526	23.173	18.257	17.125	15.074	13.036	11.544	12.251	16.910	18.036	19.579	12.424
780612 12:00	33.057	31.624	24.907	26.254	25.401	20.216	20.677	17.125	15.074	13.036	11.544	12.251	16.910	18.036	12.424
790004 12:00	25.759	25.053	19.988	20.487	20.381	18.211	17.125	15.074	13.036	11.544	12.251	16.910	18.036	19.579	12.424
790904 12:00	29.653	29.653	22.121	24.279	24.669	17.528	15.990	13.547	10.445	8.869	8.869	11.312	14.812	16.536	10.175
790915 12:00	28.641	28.115	22.146	23.144	23.144	17.528	15.990	13.547	10.445	8.869	8.869	11.312	14.812	16.536	10.175
791007 12:00	28.641	28.115	22.146	23.144	23.144	17.528	15.990	13.547	10.445	8.869	8.869	11.312	14.812	16.536	10.175
791108 12:00	32.748	29.136	22.620	23.030	23.477	17.195	15.080	13.206	9.962	8.598	8.723	10.816	13.304	15.214	9.519

7C

TIME	10	20	30	40	55	70	85	100	115	130	145	160	175	190	205
7:04:08 12:00	34,371	34,809	27,079	23,390	24,701	20,904	17,426	16,487	17,178	29,884	33,844	33,759	34,476	32,880	38,471
7:04:15 12:00	5,172	15,512	13,183	10,475	10,544	11,497	12,825	15,371	18,138	24,888	34,227	34,172	33,394	32,607	31,627
7:04:22 12:00	13,114	13,437	9,574	9,400	10,655	11,557	13,514	14,770	19,457	31,895	34,124	33,218	33,078	32,718	37,681
7:04:29 12:00	13,219	9,442	9,618	9,618	10,844	11,019	12,536	13,578	17,968	30,143	34,711	33,107	33,202	32,623	38,178
7:04:36 12:00	24,317	26,939	11,957	10,517	11,606	12,308	14,064	15,540	19,004	30,388	34,473	33,169	33,202	32,623	38,178
7:04:43 12:00	17,892	27,053	15,859	15,648	17,726	18,078	19,451	19,044	29,065	34,277	34,277	32,904	33,291	32,719	38,471
7:04:50 12:00	52,460	26,279	13,543	10,936	11,147	11,288	13,473	15,889	17,631	30,484	34,799	33,172	33,172	32,533	38,471
7:04:57 12:00	43,361	31,873	23,391	16,455	15,869	18,524	20,111	23,798	28,659	31,182	33,544	33,914	33,172	32,533	38,471
7:05:04 12:00	15,103	22,523	12,725	11,422	12,797	18,232	20,020	21,009	27,467	31,182	34,207	33,374	33,148	32,534	35,309
7:05:11 12:00	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111	16,111
7:05:18 12:00	31,973	27,004	18,131	14,195	15,130	15,130	15,130	15,130	15,130	15,130	15,130	15,130	15,130	15,130	15,130
7:05:25 12:00	18,761	22,444	12,098	10,472	11,743	13,268	16,762	16,797	25,163	32,253	34,551	33,456	33,131	32,815	37,545
7:05:32 12:00	29,387	27,708	16,306	14,433	14,645	14,142	15,981	16,797	25,163	32,253	34,551	33,456	33,131	32,815	37,545
7:05:39 12:00	31,237	25,229	14,222	12,454	14,645	14,142	15,981	16,797	25,163	32,253	34,551	33,456	33,131	32,815	37,545
7:05:46 12:00	39,786	24,997	14,117	12,703	14,082	14,753	16,168	17,547	22,148	31,514	34,551	33,456	33,131	32,815	37,545
7:05:53 12:00	31,810	23,558	15,780	13,446	14,047	14,753	16,168	17,547	22,148	31,514	34,551	33,456	33,131	32,815	37,545
7:06:00 12:00	27,132	21,773	12,649	10,724	11,296	12,528	14,245	16,048	21,368	29,219	34,551	33,456	33,131	32,815	37,545
7:06:07 12:00	26,317	22,185	12,649	10,724	11,296	12,528	14,245	16,048	21,368	29,219	34,551	33,456	33,131	32,815	37,545
7:06:14 12:00	29,381	24,997	14,438	12,210	12,245	12,316	14,276	16,276	19,070	29,219	34,551	33,456	33,131	32,815	37,545
7:06:21 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:06:28 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:06:35 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:06:42 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:06:49 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:06:56 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:03 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:10 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:17 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:24 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:31 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:38 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:45 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:52 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:07:59 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:06 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:13 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:20 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:27 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:34 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:41 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:48 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:08:55 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:02 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:09 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:16 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:23 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:30 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:37 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:44 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:51 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:09:58 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:05 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:12 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:19 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:26 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:33 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:40 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:47 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:10:54 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:11:01 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585
7:11:08 12:00	35,992	27,508	14,601	12,649	12,487	12,629	13,980	15,963	20,259	32,095	34,676	33,676	32,903	32,475	36,585

Table 7L

TIME	10	20	30	40	55	70	85	100	115	130	145	160	175	190	205
766408 12:00	35.762	27.839	17.804	15.658	14.363	13.255	10.481	8.078	8.152	7.950	7.301	8.114	10.759	5.860	7.376
766415 12:00	4.405	9.761	9.026	7.353	5.929	5.902	5.419	5.828	5.567	6.683	6.423	6.720	7.687	5.860	7.376
766704 12:00	10.158	9.498	6.416	5.520	4.886	4.811	4.811	4.736	5.222	5.371	5.184	6.229	5.731	6.534	6.014
766907 12:00	9.176	8.571	5.815	4.914	4.202	3.977	3.752	3.414	3.329	3.451	3.789	4.052	4.127	4.614	5.365
761010 12:00	12.357	11.111	7.635	6.395	5.643	5.718	6.019	5.755	6.019	6.094	5.267	5.229	5.192	5.229	5.643
770322 12:00	28.714	23.406	15.403	11.026	9.177	9.215	9.479	9.517	10.686	11.818	10.611	13.026	12.346	13.139	12.950
770419 12:00	31.043	24.563	15.679	11.375	9.160	9.638	9.624	9.185	9.940	10.091	9.034	11.450	11.299	12.847	13.565
770621 12:00	9.838	10.069	6.623	5.754	5.338	5.490	5.792	5.830	6.623	6.661	8.021	8.777	9.117	10.704	11.158
770729 12:00	26.372	20.146	13.315	10.065	8.780	8.780	8.855	8.591	9.195	9.384	7.381	8.175	8.061	9.422	10.518
770816 12:00	24.424	18.527	11.315	8.517	7.383	7.421	7.723	7.497	8.139	8.290	7.659	8.668	8.253	9.273	10.370
770828 12:00	17.434	14.719	9.464	7.082	6.533	6.591	6.817	6.742	7.006	7.233	6.855	6.956	9.275	9.124	8.670
770921 12:00	17.438	11.049	7.575	6.289	5.306	5.344	5.495	5.381	5.797	6.074	6.970	7.121	7.197	7.159	7.953
771005 12:00	22.893	16.210	10.846	8.710	7.878	7.841	7.727	6.923	6.706	6.139	5.382	6.744	7.387	7.727	8.521
771022 12:00	23.940	18.660	11.699	8.728	7.728	7.935	8.106	7.652	7.690	6.782	5.610	6.858	6.896	7.768	8.711
771121 12:00	30.198	20.967	12.759	9.771	8.561	8.561	8.901	8.372	8.409	7.766	6.140	7.199	7.237	8.031	9.015
780505 12:00	31.592	23.616	13.744	9.810	8.750	8.561	8.599	8.541	9.431	9.999	9.449	11.890	11.852	13.782	13.971
780823 12:00	14.472	11.665	7.540	6.216	5.422	5.422	5.422	5.157	5.308	5.384	5.422	6.897	7.540	9.243	9.394
781013 12:00	19.982	16.369	9.962	7.541	6.295	6.357	6.406	6.217	5.876	5.574	5.064	6.065	6.444	7.049	7.919
781101 12:00	24.146	15.814	9.244	7.201	6.255	6.330	6.293	5.990	5.498	5.309	5.158	6.368	6.330	7.692	8.033
790612 12:00	23.147	19.229	11.688	8.541	7.043	6.817	7.005	6.523	7.004	7.567	7.220	10.041	8.928	11.351	11.949
790702 12:00	10.744	9.619	7.526	5.929	5.149	5.075	5.186	5.223	5.669	5.966	6.041	8.084	8.009	9.457	9.902
790904 12:00	19.329	14.099	10.191	8.495	7.057	6.725	6.209	5.951	5.951	5.766	5.360	5.545	5.582	6.025	6.798
790915 12:00	14.312	11.919	7.778	6.567	6.055	6.091	5.835	5.431	5.951	5.504	5.174	5.652	5.652	5.871	6.441
791002 12:00	11.673	10.568	7.203	5.887	5.338	5.375	5.302	4.936	5.046	5.009	4.444	5.229	5.339	5.521	6.351
791108 12:00	21.470	15.134	9.824	8.258	7.203	6.948	6.111	14.444	4.982	4.800	4.582	5.237	5.347	5.820	6.511

Tables 9A - 9D Soil moisture contents in mm for the different intervals and periods for Skediga plots.

9A

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760407 12:00	157.18	212.90	291.75	55.722	64.63
760615 12:00	64.87	116.19	198.04	51.315	78.20
760704 12:00	70.52	111.57	191.31	41.042	76.36
760824 12:00	51.48	87.08	159.21	35.596	70.55
760907 12:00	62.40	98.96	171.83	36.562	70.10
761010 12:00	109.17	160.80	246.62	51.635	77.86
761104 12:00	126.06	178.21	263.33	52.153	75.24
761206 12:00	150.02	217.21	329.14	67.189	100.75
770208 12:00	151.31	202.77	293.30	51.462	78.62
770321 12:00	171.95	236.31	346.08	64.360	96.34
770418 12:00	161.92	231.48	348.07	69.556	104.41
770523 12:00	108.45	158.86	251.43	50.404	85.54

Table 9B

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760407 12:00	172.16	236.22	339.31	64.05	88.97
760615 12:00	93.54	163.34	256.47	69.80	85.67
760704 12:00	135.86	231.06	393.12	95.20	154.74
760824 12:00	69.11	112.06	184.08	42.94	67.76
760907 12:00	69.34	110.35	178.60	41.01	63.77
761010 12:00	91.95	145.18	218.65	53.23	65.86
761104 12:00	90.79	143.03	217.48	52.23	67.18
761206 12:00	127.23	196.53	274.55	69.30	65.45
770321 12:00	202.08	313.32	419.34	111.24	84.02
770418 12:00	196.46	310.52	456.40	114.06	128.41
770523 12:00	144.03	251.64	403.90	107.60	141.71
770620 12:00	105.27	199.86	339.75	94.98	133.52
770729 12:00	152.45	262.43	398.91	109.97	123.12
770816 12:00	146.71	252.92	393.69	106.21	128.87
770828 12:00	131.34	231.73	371.80	100.39	130.34
770920 12:00	119.24	208.23	339.87	88.98	123.52
771004 12:00	136.14	224.14	353.37	88.00	119.16
771022 12:00	135.75	220.33	344.41	84.57	113.87
771121 12:00	140.07	238.01	367.63	97.93	118.12
780505 12:00	179.52	291.04	446.27	111.52	140.82
780823 12:00	108.11	205.56	345.42	97.45	132.89
781011 12:00	140.91	235.96	370.17	95.05	123.41
781101 12:00	135.23	228.58	360.28	93.34	121.42
790612 12:00	148.41	254.71	414.89	106.29	150.15
790702 12:00	114.68	207.17	353.91	92.48	140.33
790904 12:00	139.15	230.36	354.57	91.20	112.94
790915 12:00	127.97	217.82	338.58	89.85	110.46
791002 12:00	122.59	206.46	321.68	83.86	105.54
791108 12:00	135.60	223.59	339.54	87.99	104.53

Table 9C

SUMMATION DEPTHS						
TIME	0-50	0-100	0-200	50-100	10-200	
760408 12:00	145.85	246.89	527.48	101.03	284.16	
760615 12:00	52.18	113.37	372.12	61.18	274.17	
760704 12:00	57.08	118.94	390.92	61.85	288.20	
760907 12:00	60.07	119.13	387.07	59.05	283.72	
761010 12:00	85.27	150.99	426.65	65.72	288.88	
761206 12:00	138.74	231.49	538.65	92.75	315.17	
770208 12:00	134.41	197.39	473.92	62.98	284.91	
770322 12:00	162.56	258.06	566.71	95.49	314.01	
770419 12:00	140.29	235.81	541.07	95.52	312.61	
770525 12:00	92.97	163.68	442.59	70.70	291.12	
770617 12:00	63.12	128.15	407.13	65.02	294.97	
770729 12:00	113.91	197.30	494.65	83.38	307.96	
770816 12:00	103.16	176.54	465.32	73.38	300.68	
770828 12:00	87.99	159.17	453.82	71.57	309.02	
770920 12:00	94.50	157.51	433.54	63.01	288.59	
771005 12:00	122.48	197.27	453.52	74.79	262.50	
771022 12:00	105.25	178.99	462.71	73.73	294.82	
771121 12:00	118.19	195.34	485.64	77.14	300.37	
780505 12:00	107.37	187.21	476.03	79.83	299.99	
780623 12:00	83.92	149.10	420.15	65.17	283.98	
781013 12:00	108.09	177.05	453.55	68.95	287.05	
781101 12:00	114.52	181.58	462.43	67.06	291.37	
790612 12:00	98.98	169.81	453.22	70.82	295.46	
790702 12:00	59.26	123.24	392.76	63.98	285.02	
790904 12:00	118.52	200.70	485.18	82.17	293.36	
790915 12:00	104.82	175.87	447.13	71.04	281.38	
791002 12:00	103.37	166.48	434.80	63.11	279.12	
791108 12:00	115.92	192.18	467.18	76.25	283.78	

Table 9D

SUMMATION DEPTHS						
TIME	0-50	0-100	0-200	50-100	10-200	
760408 12:00	122.45	182.07	274.81	59.619	83.26	
760615 12:00	36.19	64.97	130.37	28.778	65.44	
760704 12:00	39.27	63.36	119.26	24.094	55.10	
760907 12:00	35.34	54.75	93.64	19.407	37.21	
761010 12:00	46.68	75.66	133.35	28.977	55.78	
770322 12:00	97.95	144.61	262.44	46.652	114.99	
770419 12:00	103.44	151.30	258.54	47.861	102.56	
770621 12:00	39.97	67.94	143.69	27.970	76.57	
770729 12:00	87.79	131.66	220.72	43.871	84.53	
770816 12:00	78.97	116.54	200.96	37.570	81.01	
770828 12:00	60.82	94.18	175.05	33.361	79.45	
770921 12:00	53.97	80.89	147.51	26.928	65.09	
771005 12:00	74.27	112.67	182.37	38.401	65.13	
771022 12:00	79.56	119.05	192.38	39.492	68.48	
771121 12:00	93.37	136.55	216.94	43.174	74.43	
780505 12:00	99.19	142.30	249.52	43.102	103.50	
780823 12:00	50.04	76.95	138.79	26.913	60.24	
781013 12:00	67.38	99.73	162.02	32.353	58.31	
781101 12:00	71.84	103.09	165.02	31.247	57.56	
790612 12:00	78.08	112.57	197.45	34.483	81.94	
790702 12:00	41.96	67.70	134.00	25.747	66.15	
790904 12:00	65.66	98.35	159.22	32.687	56.89	
790915 12:00	50.89	80.42	137.72	29.532	54.85	
791002 12:00	43.97	70.36	122.70	26.392	50.43	
791108 12:00	69.28	110.21	175.31	40.928	60.31	

Tables 10A - 10D Soil moisture variations in mm
for the given periods and depth intervals, Skediga

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760615 12:00	-92.30	-96.70	-93.71	-4.406	13.572
760704 12:00	5.64	-4.62	-6.73	-10.273	-1.843
760824 12:00	-19.04	-24.48	-32.09	-5.445	-5.814
760907 12:00	10.91	11.88	12.61	0.966	-0.447
761010 12:00	46.76	61.84	74.78	15.072	7.762
761104 12:00	16.89	17.41	16.71	0.517	-2.620
761206 12:00	23.96	38.99	65.80	15.036	25.509
770208 12:00	1.28	-14.44	-35.83	-15.727	-22.131
770321 12:00	20.64	33.54	52.78	12.697	17.722
770418 12:00	-10.03	-4.83	1.99	5.196	8.065
770523 12:00	-53.46	-72.61	-96.64	-19.152	-18.870

Table 10B

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760615 12:00	-78.62	-72.87	-82.84	5.74	-3.30
760704 12:00	42.31	67.71	<u>136.65</u>	25.39	69.06
760824 12:00	-66.74	-118.99	-209.04	-52.25	-86.97
760907 12:00	0.22	-1.70	-5.47	-1.93	-3.99
761010 12:00	22.61	34.82	40.05	12.21	2.09
761104 12:00	-1.15	-2.15	-1.17	-0.99	1.31
761206 12:00	36.43	53.50	57.07	17.07	-1.73
770321 12:00	74.84	116.78	144.78	41.94	18.57
770418 12:00	-5.61	-2.80	37.06	2.81	44.39
770523 12:00	-52.42	-58.88	-52.50	-6.45	13.29
770620 12:00	-38.75	-51.77	-64.14	-13.01	-6.18
770729 12:00	47.17	62.56	59.15	15.38	-10.40
770816 12:00	-5.74	-9.50	-5.22	-3.76	5.74
770828 12:00	-15.36	-21.18	-21.89	-5.82	1.46
770920 12:00	-12.09	-23.50	-31.92	-11.40	-6.82
771004 12:00	16.90	15.91	13.50	-0.98	-4.35
771022 12:00	-0.38	-3.81	-6.96	-3.42	-5.29
771121 12:00	4.32	17.67	23.21	13.35	4.25
790505 12:00	39.44	53.03	78.64	13.59	22.70
780823 12:00	-71.40	-85.48	-100.85	-14.07	-7.93
781011 12:00	32.80	30.40	24.75	-2.40	-9.47
781101 12:00	-5.67	-7.38	-9.88	-1.70	-1.98
790612 12:00	13.17	26.12	54.60	12.95	28.72
790702 12:00	-33.72	-47.54	-60.97	-13.81	-9.81
790904 12:00	24.47	23.19	0.66	-1.27	-27.38
790915 12:00	-11.18	-12.53	-15.98	-1.35	-2.48
791002 12:00	-5.38	-11.36	-16.90	-5.98	-4.91
791108 12:00	13.01	17.13	17.85	4.12	-1.01

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760615 12:00	-93.67	-133.52	-155.35	-39.85	-9.98
760704 12:00	4.90	5.57	18.79	0.67	14.02
760907 12:00	2.98	0.18	-3.84	-2.79	-4.48
761010 12:00	25.19	31.86	39.58	6.66	5.15
761206 12:00	53.47	80.50	111.99	27.02	26.29
770208 12:00	-4.32	-34.10	-64.73	-29.77	-30.26
770322 12:00	28.15	60.66	92.79	32.51	29.09
770419 12:00	-22.27	-22.24	-25.64	0.02	-1.39
770525 12:00	-47.32	-72.13	-98.48	-24.81	-21.49
770617 12:00	-29.84	-35.52	-35.45	-5.68	3.84
770729 12:00	50.79	69.15	87.52	18.35	12.98
770816 12:00	-10.75	-20.76	-29.33	-10.00	-7.28
770828 12:00	-15.56	-17.37	-11.49	-1.80	8.34
770920 12:00	6.90	-1.65	-20.28	-6.56	-20.42
771005 12:00	27.98	39.76	19.98	11.77	-26.09
771022 12:00	-17.23	-18.28	9.18	-1.05	32.32
771121 12:00	12.93	16.34	22.93	3.40	5.55
780505 12:00	-10.82	-8.12	-9.61	2.69	-0.78
780823 12:00	-23.44	-38.11	-55.87	-14.66	-15.60
781013 12:00	24.17	27.94	33.40	3.77	3.06
781101 12:00	6.42	4.53	8.87	-1.88	4.31
790612 12:00	-15.53	-11.77	-9.20	3.75	4.09
790702 12:00	-39.72	-46.56	-60.46	-6.83	-10.44
790904 12:00	59.25	77.45	92.42	18.19	8.34
790915 12:00	-13.70	-24.82	-38.04	-11.12	-11.98
791002 12:00	-1.45	-9.38	-12.32	-7.93	-2.25
791108 12:00	12.55	25.69	32.37	13.14	4.66

Table 10D

SUMMATION DEPTHS					
TIME	0-50	0-100	0-200	50-100	10-200
760615 12:00	-86.25	-117.10	-144.43	-30.840	-17.81
760704 12:00	3.08	-1.60	-11.10	-4.684	-10.34
760907 12:00	-3.93	-6.61	-25.61	-4.687	-17.89
761010 12:00	11.34	20.91	39.70	9.569	18.56
770322 12:00	51.27	68.94	129.09	17.675	59.21
770419 12:00	5.48	6.69	-3.90	1.209	-12.43
770621 12:00	-63.46	-83.36	-114.85	-19.891	-25.98
770729 12:00	47.81	63.72	77.02	15.901	7.95
770816 12:00	-8.82	-15.12	-19.75	-6.301	-3.51
770828 12:00	-18.14	-22.35	-25.91	-4.208	-1.56
770921 12:00	-6.85	-13.28	-27.54	-6.432	-14.35
771005 12:00	20.30	31.77	34.86	11.472	0.04
771022 12:00	5.29	6.38	10.01	1.091	3.34
771121 12:00	13.81	17.49	24.55	3.681	5.95
780505 12:00	5.82	5.74	32.57	-0.072	29.06
780823 12:00	-49.15	-65.34	-110.72	-16.188	-43.25
781013 12:00	17.34	22.78	23.22	5.439	-1.92
781101 12:00	4.46	3.35	2.99	-1.105	-0.75
790612 12:00	6.24	9.47	32.42	3.235	24.37
790702 12:00	-36.12	-44.86	-63.44	-8.736	-15.79
790904 12:00	23.70	30.64	25.21	6.939	-9.25
790915 12:00	-14.77	-17.93	-21.49	-3.154	-2.04
791002 12:00	-6.91	-10.05	-15.01	-3.140	-4.42
791108 12:00	25.31	39.85	52.60	14.536	9.88

Table 11. Soil moisture content by volume % for plot C in Tärnsjö.

DATE	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150	160	170	180	190	200
770118	19.2	16.6	14.2	15.9	14.8	13.7	13.5	13.6	13.5	16.2	15.9	15.3	11.2	8.8	16.2	14.9	17.9	35.6	21.1	
770502	13.8	14.2	11.9	12.1	12.3	11.8	10.8	10.4	12.6	12.4	10.8	15.2	10.5	8.3	15.7	17.8	23.2	26.2	22.5	
770516	13.7	13.0	11.8	12.8	12.3	9.5	9.7	9.5	7.3	7.8	10.1	14.3	9.8	7.6	15.0	11.9	17.0	24.2	22.5	
770528	4.2	9.8	9.8	11.1	9.5	9.3	9.4	9.6	6.8	7.3	9.3	13.6	9.2	6.7	12.9	11.5	15.1	22.9	20.0	
770617	7.0	8.3	8.3	8.5	7.0	8.5	7.4	8.5	7.0	6.3	7.2	8.6	12.5	8.6	5.1	12.9	13.2	14.1	21.2	19.3
770704	8.8	10.5	10.7	11.9	10.7	10.0	10.7	11.1	7.8	8.5	10.0	14.7	12.6	9.5	17.0	15.3	15.3	22.2	23.8	
770815	11.8	11.2	10.3	12.3	10.2	8.8	9.8	9.9	9.2	6.5	9.1	13.4	9.7	6.7	14.0	13.5	14.0	22.6	23.8	
770915	11.4	11.2	10.7	12.2	11.0	9.5	8.2	9.0	7.3	6.9	7.7	12.7	9.4	5.5	12.2	11.9	13.8	19.9	20.5	
771017	12.3	12.2	11.6	12.3	16.8	8.8	10.5	9.5	7.6	8.2	10.0	15.6	8.6	8.5	15.4	12.5	15.4	25.3	21.6	
771116	18.2	16.1	15.0	16.1	14.1	12.2	12.3	12.8	9.6	9.5	10.8	16.1	14.6	8.1	17.3	15.9	19.2	28.1	24.1	
771216	16.0	12.8	11.1	11.4	10.0	9.3	11.6	11.8	9.2	7.5	9.5	9.5	13.7	9.5	6.5	14.0	13.5	15.8	24.3	20.8

DATE	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150	160	170	180	190	200
780116	23.4	17.0	13.6	13.0	10.7	8.9	11.6	8.9	5.8	7.7	10.0	13.7	8.9	6.2	13.6	13.0	13.5	22.6	20.1	
780216	22.7	15.0	12.3	13.8	11.3	6.9	7.0	7.0	5.6	5.7	6.4	11.5	5.4	3.8	11.3	8.3	13.5	14.1	15.6	
780316	25.1	12.7	14.1	15.7	12.7	10.3	9.9	8.3	6.8	7.2	7.3	13.2	7.2	5.8	11.9	12.3	16.3	23.8	17.0	
780328	23.4	18.7	19.1	17.7	10.7	7.4	9.4	8.0	7.6	8.5	12.4	12.1	6.8	6.8	12.2	13.3	19.7	23.4	18.1	
780406	18.8	16.6	17.3	22.7	12.8	13.0	13.5	10.5	10.3	9.5	10.3	9.5	11.0	10.0	10.8	18.4	18.4	21.4	27.8	
780417	16.8	16.0	14.8	17.0	15.8	12.4	11.8	12.5	9.3	8.9	10.8	15.2	11.0	8.1	16.2	15.7	18.8	26.5	23.3	
780428	15.4	17.1	13.5	15.3	13.8	11.0	11.3	10.8	8.3	9.5	11.1	14.6	8.9	7.3	14.6	12.5	18.4	25.4	22.9	
780508	12.0	12.0	10.7	12.8	10.2	9.3	10.4	9.5	5.8	7.7	9.9	14.8	8.1	6.7	14.0	11.5	16.2	25.0	21.0	
780518	14.8	13.2	10.7	12.3	10.7	9.5	9.9	9.3	7.3	8.2	9.7	14.1	10.1	6.7	14.0	11.5	16.2	24.2	22.4	
780529	9.1	10.1	9.3	9.8	8.3	7.4	9.7	9.0	5.8	8.2	11.8	13.6	5.4	6.7	12.8	14.4	17.7	25.0	17.2	
780608	4.1	6.4	6.7	7.1	5.1	4.1	5.7	6.2	4.0	5.6	7.1	11.8	10.3	5.6	13.2	12.3	13.5	21.3	22.5	
780619	7.8	8.3	7.9	8.5	6.6	6.2	7.8	6.7	5.6	6.6	10.2	13.0	8.6	6.7	12.5	12.5	18.5	27.8	15.6	
780717	7.8	9.0	10.0	11.1	8.9	8.8	11.6	8.9	7.1	7.7	9.9	13.0	8.6	5.8	14.0	13.0	13.5	24.6	20.7	
780817	7.7	9.3	9.7	11.2	9.1	8.2	9.0	8.6	6.9	7.5	9.5	14.0	8.1	6.2	13.9	13.5	16.0	24.4	18.6	
780918	10.1	11.2	11.4	13.0	10.8	9.9	9.9	8.9	7.3	7.7	9.5	14.2	9.9	7.8	13.0	14.5	16.6	24.4	22.1	
781016	9.3	9.8	9.9	10.4	10.7	8.4	11.6	9.0	6.9	7.2	9.0	13.7	7.7	5.6	13.9	12.8	15.3	23.2	19.9	
781115	10.9	10.3	12.6	12.8	11.0	9.5	11.6	9.4	7.1	7.0	9.0	13.0	7.6	5.3	12.8	11.9	12.5	23.2	19.9	
781215	14.3	11.2	10.7	12.5	11.0	9.5	10.4	9.5	8.5	8.5	11.2	14.1	6.5	6.2	12.8	13.2	17.0	23.1	17.3	

DATE	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150	160	170	180	190	200
790116	14.8	12.6	12.9	15.1	12.5	9.5	9.5	9.0	8.5	7.6	7.7	10.2	13.6	7.6	5.8	13.1	13.5	15.8	21.9	18.1
790215	15.2	13.2	13.7	15.3	13.4	12.5	12.5	8.8	8.0	6.9	8.0	10.2	14.2	7.2	5.3	12.4	12.3	16.3	24.2	17.3
790301	16.3	14.1	13.6	16.4	13.8	14.0	11.3	10.3	8.3	6.8	8.0	10.2	14.2	9.6	5.8	12.4	13.0	15.5	23.2	19.8
790316	21.6	21.4	19.0	20.1	13.4	11.1	10.7	10.2	8.3	8.0	10.0	14.0	7.0	4.9	11.9	11.9	14.8	23.1	17.9	
790402	30.7	29.8	26.2	24.6	16.0	13.8	12.4	12.4	5.8	9.7	11.5	16.1	9.5	6.2	13.9	12.3	14.6	22.6	16.3	
790418	16.3	15.8	15.8	15.9	13.4	11.7	12.4	12.1	9.3	9.3	11.2	15.6	11.0	9.9	18.4	16.7	19.2	25.4	24.7	
790503	16.0	14.8	14.5	13.9	13.5	11.7	11.6	11.1	9.5	9.7	11.2	14.7	10.0	8.8	16.6	15.3	19.5	25.7	23.5	
790603	12.7	12.0	11.1	12.3	10.7	9.5	9.9	10.6	8.3	8.5	10.7	14.8	8.6	7.3	14.9	14.9	17.9	28.0	21.7	
790615	7.8	12.5	9.6	12.8	12.7	11.2	10.6	10.8	2.6	9.7	10.2	13.6	9.6	6.5	14.4	13.8	13.3	23.1	22.1	
790618	6.8	6.8	6.8	9.7	8.3	6.8	6.8	6.8	5.7	5.7	10.7	13.1	8.2	5.6	13.6	13.8	16.8	24.1	20.1	
790716	5.8	6.8	6.8	9.7	8.3	6.8	6.8	6.8	5.7	5.7	10.7	13.1	8.2	5.6	13.6	13.8	16.8	24.1	20.1	
790816	3.3	7.7	7.1	8.6	6.8	6.2	6.2	6.2	5.9	6.4	7.0	9.9	7.2	6.1	6.7	7.2	9.5	14.8	23.2	
790916	8.8	8.2	7.9	8.4	7.1	7.2	8.5	7.6	6.1	6.8	8.1	14.0	8.2	6.1	12.5	12.6	16.1	24.1	18.6	

Table 12 Soil moisture contents by vol% for the different periods and depths (Tärnsjö)

PERIOD	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150	160	170	180	190	200
770806-																				
770903	11.80	11.20	10.30	12.30	10.20	8.80	9.80	9.80	9.90	9.20	6.50	9.10	13.40	9.70	6.70	14.00	13.50	14.00	22.60	23.80
770903-																				
771009	11.40	11.20	10.70	12.20	11.00	9.50	8.20	9.00	6.30	6.90	7.70	12.70	9.40	5.50	12.20	11.90	13.8	19.90	20.50	
770806-																				
771009	11.83	11.53	10.87	12.27	12.67	9.03	4.50	9.47	7.70	7.20	8.93	13.90	9.23	6.90	13.87	12.63	14.40	22.60	21.97	
780512-																				
780609	9.33	9.90	8.90	9.73	8.03	7.00	8.43	8.23	5.70	7.33	9.53	13.17	8.60	6.33	13.33	12.73	15.83	23.50	20.70	
780609-																				
780905	9.85	9.65	9.75	10.95	8.50	8.25	9.57	8.28	6.72	7.38	9.78	13.55	8.05	6.58	13.35	13.38	16.15	25.30	19.30	
780512-																				
780905	8.77	4.76	9.39	10.43	7.20	7.73	9.04	8.26	6.29	7.36	9.67	13.39	8.29	6.47	13.34	13.10	16.01	24.43	14.74	
790904-																				
791006	5.30	7.70	7.10	8.60	6.80	6.20	6.10	5.90	6.40	7.50	6.80	7.40	7.10	5.40	7.20	9.20	9.30	9.90	9.70	
771006-																				
771109	8.80	8.20	7.90	8.40	7.10	7.20	8.50	7.60	6.10	6.80	8.10	14.00	8.20	6.10	12.50	12.60	16.10	24.10	14.40	
770904-																				
771109	6.05	7.95	7.50	8.50	6.95	6.60	7.30	6.75	6.25	7.15	7.45	10.70	7.65	5.75	9.85	10.90	12.70	17.00	14.15	

ACKNOWLEDGMENTS

This investigation is financially supported by the Swedish Natural Science Research Council (NFR) and carried out at the Division of Hydrology in Uppsala. The computer programs for data processing were made available by P. E. Jansson who also assisted in doing the computations. The author is grateful to prof. Erik Eriksson who sponsored the project and whose encouragement has been indispensable. The moral support and the assistance rendered by my wife in typing and correcting the manuscript is a major contribution towards the completion of this work. I finally wish to express my indebtedness to all those at the division who kindly read and gave many constructive comments on the manuscript.

REFERENCES

- [1] Allison, G.B. and Hughes, M.W. 1974. "Environmental tritium in the unsaturated zone. Estimation of recharge to an unconfined aquifer", Isotope Techniques in Groundwater Hydrology (Proc. Symp. Vienna, 1974), IAEA, Vienna. 1, 58 - 72.
- [2] Andersen, L.J. and Sevel, T. 1974. "Six years environmental tritium profiles in the unsaturated zones", Ibidem, 3 - 17.
- [3] Athavale, R.N., Murti, G.S. and Chand, R. 1980. Estimation of recharge to the phreatic aquifers of the Lower Maner Basin, India, by using the tritium injection method. Journal of Hydrology, 45(1980) 185 - 202.
- [4] van Bavel, C.H.M., Nielsen, D.R. and Davidson, J.M. 1961. Calibration and characteristics of two neutron probes. Proc. Soil. Sci. Am. 25, 329 - 333.
- [5] Baver, L.D. 1956. "Soil physics." Wiley, New York.
- [6] Beven, K. 1979. A sensitivity analysis of the Penman-Monteith actual evapotranspiration estimates, Journal of Hydrology, 44(1979) 169 - 190.
- [7] Blume, H.P., Zimmermann, U. and Munnich, K.O. 1967. "Tritium tagging of soil moisture: the water balance of forest soils", Isotope and Radiation Techniques in Soil Physics and Irrigation Studies. Proc., IAEA Symp. Vienna, 315 - 332.
- [8] Brown, R.M. and Grummitt, W.E. 1956. The determination of tritium in natural waters, Canadian J. of Chem., 34, 220 - 226.
- [9] Cihdar, J. and Ulaby, F.T. 1974. Dielectric properties of soils as a function of moisture content, RSL Tech. Rep. No. 177 - 47, Univ. of Kansas Center for Research.
- [11] Childs, E.C. 1945. The water table equipotentials and stream lines in drained land, Soil Sci. 59, 405.

- [12] Corey, J.C. and Nielsen, D.R. 1973. "Some factors influencing the interpretation of soil-solute-water interactions", Soil Moisture and Irrigation Studies, IAEA Proc. Panel. Vienna, 2, 87 - 104.
- [13] Croney, D., Coleman, J.D. and Currer, E.W.H. 1951. The electrical resistance method of measuring soil moisture. British J. Appl. Physics, 2, 85 - 91.
- [14] Datta, P.S. and Goel, P.S. 1977. Groundwater recharge in the Panjab State, India using tritium tracers, Nordic Hydrology, 8(1977) 225 - 236.
- [15] Dressie, Z and Kifumba, F. 1977. A method for estimating groundwater recharge by using injected tritium profiles in the unsaturated zone, Vannet i Norden, 2, 43 - 49.
- [16] Dressie, Z. 1982. A new soil moisture sampling method for radio- and stable isotope analysis by sucking soil air with molecular sieve, Vannet i Norden, 3, 5 - 28.
- [17] Dunne, T. and Leopold, L.B. 1978. "Water in environmental planning" Freeman & Co. San Francisco.
- [18] Encyclopedia Britanica, 1968. 10, 949 - 951
- [19] Encyclopedia Britanica. 1968. 11, 959 - 961.
- [20] Eriksson, E. 1958. The possible use of tritium for estimating groundwater recharge, Tellus, 10, 472 - 478.
- [21] Eriksson, E. 1961. Natural reservoir and their characteristics, Geofisica International, 1, 2, 27 - 43.
- [22] Eriksson, E. 1963. Atmospheric tritium as a tool for the study of certain hydrologic aspects of river basins, Tellus, 15, 302 - 208.
- [23] Eriksson, E. 1967. "Isotopes in hydrometeorology", In, Isotopes in Hydrology, (Proc. Symp. Vienna, 1966), 21 - 32.
- [24] Eriksson, E. and Johansson, S. 1978. Gotlands grundvatten reservoir, Int. Rep.

- [25] Eriksson, E. 1973. "Nuclear techniques in hydrological studies",
In, Use of Isotopes and Radiation in Agriculture,
Biology and Animal Sciences, Symp. Proc. 51 - 54
- [25] Eriksson, E. 1975. "Hydrologic modelling" In, Procedures and Exam-
ples in Integrated Ecosystem Research, Tech. Rep. 1.
- [26] Florkowski, T. 1981. "Low level tritium assay in water samples
by electrolytic enrichment and liquid scintillation
counting in the IAEA laboratory", In, Methods of
Low-Level Counting and Spectrometry, (Proc. Symp.
Berlin(West), 1981), Vienna, 335 - 358.
- [27] Freeze, R.A. and Cherry, A.J. 1979. "Groundwater", pp.41 -44.
- [28] Gardner, W.R. 1958. Some steady-state solutions of the unsaturated
moisture flow equations with applications to evapor-
ation from a water table, Soil Sci. 85, 288 - 332.
- [29] Gasper, E. and Oncescu, M. 1972. Radioactive tracers in Hydrology:
Development in Hydrology, Vol. 1 IAEA, Vienna.
- [30] Ghuman, B.S. and Prihar, S.S. 1980. Chloride displacement by wa-
ter in layered soil column, Aust. J. of Soil Resear-
ch, 18, 2, 207 - 214
- [31] Gustafsson, G. 1978. Studies of the hydrogeology of subaqueous
eskers, Publ. A 26, Disertation, Chalmers Univer-
sity of Technology, Gothenburg.
- [32] Henriksson, A. and Johnsson, A. 1971 Infiltrationsanläggningars
ekonomi, Bulletin Ser. B 17, Lund.
- [33] Hillel, D. 1971. "Soil and water: Physical principles and processes"
Academic Press, New York.
- [34] Hillel, D. 1980 "Application of soil science"
- [35] International Atomic Energy Agency, 1968. Guide book on nuclear
techniques in hydrology, Tech. Rep. Ser. No. 91,
Vienna.

- [36] International Atomic Energy Agency. 1983. Guide book on nuclear techniques in hydrology, Tech. Rep. Ser. No. 91, Vienna.
- [37] Jansson, P.E. 1977 Jordmåns- och markbeskrivning for Inventjä-rnsheden, Swe. Con. For. Proj., Int. Rep. No. 54.
- [38] Johnson Division UOP, Inc. 1972 Groundwater and wells: a reference book for the water well industry, Saint Paul, Minnesota.
- [39] Knutsson, G. 1966 Tracers for groundwater investigation, Proc. of the Int. Symp., Stockholm.
- [40] Kreutzer, K., Strebel, O. and Renger, M. 1980 Field measurement of seepage and evapotranspiration rate for a soil under plant cover: A comparison of soil water balance and tritium labeling procedure, Journal of Hydrology, 48(1980), 137 - 146.
- [41] Linsley, K.R. Jr., Kohler, M.A. and paulhus, J.L.H. 1958 "Hydrology for engeners", McGraw-Hill Book Company, Inc., New York
- [42] Long, I:F: and French, B.K. 1967 Measurement of soil moisture in the field by neutron moderation, Journal of Soil Sci., 18, 1, 149 - 165.
- [43] Lundien, J.R. 1971 Terrain analysis by electromagnetic means. Mississippi Tech. Rep. No. 3-693 U.S. Army Engineer Waterways Exp. Stn., Vicksburg.
- [44] Marshal, T.J. and Holmes, J.W. 1979 "Soil physics"
- [45] Milanov, T. 1973 Hydrological studies in representative basin Kassjöän, Soil Water Studies, IHD Rep., 2, 37.
- [46] Milanov, T 1975 Hydrological studies in representative basin Lapträsket, Soil Water Studies, IHD Rep., 5, 46.
- [47] Norberg, L. and Persson, G. 1974 The national groundwater network of Sweden, Geological Survey of Sweden (SGU), Ser. Ca No. 48, Stockholm.

- [48] Okrasinski, T.A., Koerner, R.M. and Lord, A.E. 1978 Dielectric constant determination of soil at L band microwave frequencies, *Geotechnical Testing Journal* GTJODJ 1, 134 - 140.
- [49] Penman, H.L. 1948 Natural evaporation from open water, bare soil and grass, *Proc. Roy. Sci. (London) A*, 193.
- [50] Philip, J.R. 1958 The theory of infiltration. 6, Effect of water depth over soil, *Soil Sci.*, 85, 288 - 286.
- [51] Ibidem 1969 Theory of infiltration in the hydrologic cycle. *Adv. Hydrosience*, 5, 215 - 296.
- [52] Poulovassilis, A. 1969 A steady-state potential and moisture profiles in layered porous media, *Journal of Soil Sci.*, 107, 20, 47 - 52.
- [53] Ibidem 1974 Soil water properties of a layered soil determined insitu, In, *Isotope and Radiation Technique in Soil Physics and Irrigation Studies*, Proc., IAEA Symp., Vienna (1973), 205 - 224.
- [54] Richards, L.A. 1949b Methods for mounting porous plates used in soil moisture measurements, *Agron. J.*, 41, 487 - 490.
- [55] Richards, L.A. and Wadleigh, C.H. 1952 Soil water and plant growth Chapter 3 In, *Soil Physical conditions and Plant Growth*.
- [56] Richards, L.A. and Weeks, L.V. 1953 Capillary conductivity values from moist fields and tension measurements of soil column, *S.S., S. Am. Pr.*, 17.
- [57] Rodda, J.C., Downing, R.A. and Frank, M.L. 1976 "Systematic hydrology", *Newnes-Butterworths*, London.
- [58] Rose, C.W., Steen, W.R. and Drumond, J.E. 1965 Detrmination of hydraulic conductivity as a function of depth and water content for soil insitu, *Aust. Soil Res.*, 3, 1.

- [59] Sidenvall, J. 1970 Grundvatten i uppsalatrakten, Kvartärgeologiska institutionen, Uppsala universitet. Int. Rep.
- [60] Sidenvall, J. 1975 Safeguarding of the water supplies in Uppsala, Sweden, (Uppsala Commune), Int. Rep.
- [61] Shaw, B. and Baver, L.D. 1939b An electrothermal method for following moisture changes of soil insitu, Proc. Soil Sci. Soc. Am., 4, 78 - 83.
- [62] Smith, D.B. and Wearn, P.L. 1970 "Water movement in the unsaturated zone of low and high permeability strata by measuring natural tritium", Isotope Hydrology (Proc. Symp. Vienna, 1970) IAEA, Vienna. 73 - 88.
- [63] Sukhija, B.S. 1972 Evaluation of groundwater recharge in semi-arid regeons of India using environmental tritium, Institute of Fundamental research, University of Bombay, Disertation paper.
- [64] Taylor, C.B. 1978 Interlaboratory comparison of low-level tritium measurements i water, Int. J. Appl. Radiat. Isot., 29, 39 - 48
- [65] Theodorsson, P. 1974 Improved tritium counting through high electrolytic enrichment, Int. J. Appl. Radiat. Isot., 25, 97 - 104.
- [66] Thoma, G., Esser, N., Sontag, C., Weiss, J. and Rudolph, J. 1978 New technique of insitu soil moisture sampling for environmental isotope analysis applied at "Pilat Dune" near Bordeaux. HETP modelling of bomb tritium propagation in the unsaturated and saturated zones, In, International Symposium on Isotope Hydrology, Neuherberg.
- [67] Thomas, A.M. 1966 Insitu measurement of moisture in soil and similar substances by fringe capacitance, J. Sci. Instrum., 43, 21 - 27.

- [68] Todd, D.K. 1959 "Ground water hydrology", Willy & Sons, Inc., New York.
- [69] Topp, G.C., Davis, J.L. and Annan, A.P. 1980a Electromagnetic determination of soil water content, measurement in coaxial transmission lines, Water Resour. Res. 16, 575 - 582.
- [70] Topp, G.C. et al 1980b Electromagnetic determination of soil water content: application of TDR to field measurements, Proc. Third Colloc. on Planetary Water, University of New York, Buffalo.
- [71] Viessman, W. Jr., Knapp, J.L. and Harbaugh, T.E. "Introduction to hydrology",
- [72] Vogel, J.C., Thilo, L. and van Dijken, 1974 Determination of groundwater recharge with tritium, Journal of Hydrology, 23, 131 - 140.
- [73] Wallén, C.C. 1966 Global solar radiation and evapotranspiration in Sweden, Tellus 18, 4.
- [74] Watson, K.K. 1966 An instantaneous profile method for determining the hydraulic conductivity of unsaturated porous materials, Water Resources, 2, 709.
- [75] Whitney, M., Gardner, F.D. and Briggs, L.J. 1898 An electrical method of determining the moisture content of stable soils
- [76] Wison, E.M. 1969 "Engineering hydrology", Macmillan.
- [77] Zimmermann, U., Munich, K.O. and Roether, W. 1967 Downward movement of soil moisture traced by means of hydrogen isotopes, Geog. Monogr. Ser. 11.
- [78] Zimmermann, U., Ehalt, D. and Munich, K.O. 1966 "Soil water movement and evapotranspiration: changes in the isotopic composition of the water", In, Isotopes in Hydrology, Proc. Symp. Vienna (1966), 567 - 586.

- [79] Östlund, H.G. and Werner, E. 1961 "The electrolytic enrichment of tritium and deuterium for natural tritium measurements", In, Tritium in the Physical and Biological Sciences, 1, 95 - 104.
- [80] Östlund, H.G. and Dorsey, H.G. 1975 Rapid electrolytic enrichment and hydrogen gas proportional counting of tritium", In, Low Radioactivity measurements and Applications, Proc. Symp. The high Tatras, Czech.

PART II

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi

Zegeye Dressie

Comparison of a Numerical Solution of the Soil Water Flow and Recharge Estimates With the Tracer and the Water Balance Methods

ABSTRACT

In this experiment, the SOIL model (a numerical model which solves the physical equations for water and heat flows in the unsaturated zone by the finite difference approximations) was applied to study the short term fast infiltration process and the long term soil water flows including the temporal and spatial variations of soil moisture contents in the soil profile. Meteorological data were used as the driving variables while the water retention curves and an estimated unsaturated conductivity function were used as the soil physical properties.

The simulation method provided the possibility of predicting recharge for a longer time period in order to be able to study its temporal and vertical pattern. The long term simulation demonstrated the general behaviour of recharge processes. The regression analysis performed proved that when simulation is done for longer periods, cumulative recharge mainly depends upon the meteorological conditions (precipitation and evapotranspiration) and to a small extent on the soil physical properties. A linear relationship between recharge and precipitation can be obtained from long term simulations, indicating that recharge for similar areas can be estimated if the meteorological data is available.

For short term fast infiltration processes, the initial condition and the soil physical properties proved to be very important.

During long term simulations with low flow rates, the agreement between the simulated and observed (using the neutron scattering, the TDR (Time Domain Reflectometry), and the gravimetric methods) soil moisture contents is quite good. Anomalies such as fingering effects occur at saturated conditions when the soil physical properties are very important which also means that the the retention curve determined in the laboratory does not fully express the real field conditions.

A water model may be used if the temporal and spatial patterns are of interest.

TABLE OF CONTENTS

1	ACKNOWLEDGMENTS	7
2	LIST OF SYMBOLS	9
3	INTRODUCTION	11
4	SCOPE OF THE STUDY	13
5	NUMERICAL SIMULATION - THE SOIL MODEL	14
6	AREAS OF INVESTIGATION	20
6.1	Area 1 (Uppsala site)	21
6.2	Area 2 (Skediga site)	22
7	EXPERIMENTAL PROCEDURE	23
7.1	Infiltration scheme	23
7.2	Laboratory work	26
8	FIELD DATA - RESULTS AND DISCUSSION	27
8.1	Saturated conductivity	27
8.2	Tritium profiles	28
8.3	Soil moisture measurements	29
9	INFILTRATION PROCESS SIMULATION	32
9.1	Macropores and soil water flow	32
9.2	Parameters	36
9.2.1	Plot A (Uppsala site)	36
9.2.1.1	The soil characteristic curves	36
9.2.1.2	Meteorological data	39
9.2.2	Plot 2 (Skediga site)	44
9.2.2.1	The soil characteristic curves	44
9.2.2.2	Meteorological data	45
9.3	High intensity infiltration process - Results and discussion	45
10	LONG TERM SIMULATIONS AND RECHARGE ESTIMATES	57
10.1	Simulations	57
10.2	Results: Plot A - Period 1	58
10.3	Results: Plot A - Period 2	62
10.4	Results: Plot B - Period 1	65
10.5	Results: Plot B - Period 2	70
10.6	Discussions	77
10.6.1	Uppsala site	77
10.6.2	Skediga site	78
11	STATISTICAL ANALYSIS	83
12	CONCLUSIONS	87
13	REFERENCES	89

1 ACKNOWLEDGMENTS

The author expresses his sincere gratitude to Dr Per Erik Jansson from the Swedish University of Agricultural Sciences, Department of Soil Sciences, who, not only provided with the simulation code, but was always available for intensive discussions and guidance concerning the implementation and interpretation of the simulation work. Professor Erik Erikssons guidance and moral support was indispensable. Professor Lars Bengtsson critically read the manuscript and gave many constructive comments. Thanks are also due to my colleagues for the stimulating discussions and for reading and commenting on the paper.

2 LIST OF SYMBOLS

c_p	=	Specific heat	$\text{j}/(\text{kg} \cdot ^\circ\text{C})$
e	=	Vapour pressure	Pa
g	=	Gravitational force	cm/sec^2
K	=	Hydraulic conductivity	cm/sec
K_s	=	Saturated hydraulic conductivity	cm/sec
L	=	Total length travelled by water particle	cm
L_e	=	Effective (shortest) path the water particle takes ..	cm
L_v	=	Latent heat of vaporization	j/kg
n	=	Tortuosity factor (L_e/L)	none
p	=	Porosity	vol%
θ	=	Moisture content	vol%
θ_r	=	Residual soil moisture content	vol%
θ_s	=	Saturated soil moisture content	vol%
q	=	Soil water flux	$\text{kg}/(\text{m}^2 \cdot \text{sec})$
q_h^s	=	Heat flow into soil	$\text{j}/(\text{m}^2 \cdot \text{sec})$
r_a	=	Aerodynamic resistance	sec/cm
r_s	=	Surface resistance	sec/cm
R_{is}	=	Global radiation	$\text{j}/(\text{m}^2 \cdot \text{sec})$
R_n	=	Net radiation	$\text{j}/(\text{m}^2 \cdot \text{sec})$
s	=	Source/sink term	$\text{kg}/(\text{m}^2 \cdot \text{sec})$
S_e	=	Effective saturation	none
t	=	Time	sec
T_a	=	Air temperature	$^\circ\text{C}$
TR_p	=	Potential transpiration	$\text{kg}/(\text{m}^2 \cdot \text{day})$
u	=	Wind speed	m/sec

ψ	=	Matric potential (capillary pressure)	cm
ψ_a	=	The tention at which air bubbles start to enter into the soil voids	cm
λ	=	A parameter that reflects the pore size distribution of the soil	none
γ	=	Psychrometric constant	Pa/ $^{\circ}$ C
κ	=	von Karmans constant	none
ρ	=	Density of air	kg/m ³
Δ	=	The slope of a saturated vapour prussure versus tempera- ture curve	Pa/ $^{\circ}$ C
z	=	Depth	cm
K_r	=	Relative conductivity of soil	cm/sec
δe	=	Vapour pressure deficit	Pa

3 INTRODUCTION

Most of the water falling as rain or/and snow on the surface of the earth, passes through the unsaturated soil zone during the subsequent processes of infiltration, redistribution, evaporation/transpiration and recharge to ground water (J. R. Philip, 1969). Hence, such problems involving ground water recharge, irrigation, drainage, water conservation, pollution including infiltration-runoff control, require a thorough understanding of the patterns of water movement in the aerated soil zone (D. Hillel et al, 1972). Therefore the process of soil moisture movement in the unsaturated soil profile is of prime importance not only to the hydrologist but also to the agriculturalist, the ecologist and the engineer. A great deal of pioneering research work on soil water movement and capillary suction has been carried out by several investigators a review of which is made by Seckel (1978).

This work is a continuation of the previous study which deals with the estimation of ground water recharge using tracers to follow the vertical flow of soil moisture. In the earlier experiment (Dressie, 1984), studies of the temporal changes of soil moisture contents and recharge formation were made using the tritium-tagging and the water balance methods. The injected tritium method was further coupled with the stable isotope ($O - 18$) method and recharge estimates were done for a selected area near Uppsala (Saxena and Dressie, 1984). The soil moisture contents were monitored during a period of 4 years (1976 - 79) with intervals of one week to one month (except during the winter months) using the neutron scattering method and were compared with those obtained by the gravimetric method.

To obtain a general knowledge from such experiments, a detailed analysis of the mechanisms and patterns is needed. This is of special importance if the estimated ground water recharge is to be extended to new time periods or to other similar areas. Physically based numerical model may be used as a useful tool for interpreting and for generalization of experimental

results. That is why many attempts have been made to use numerical or analytical models (van Keulen and van Beek, 1972; Kastanek, 1971) and analog models (Wind, 1972).

In this study, the SOIL model (Jansson and Halldin, 1980) based on the numerical solutions by finite difference approximations of the differential equations of water and heat flows, was used to estimate ground water recharge. The results thus obtained were compared with those previous ones estimated by the tracer and the water balance methods.

The spatial and temporal variations of simulated and measured soil moisture contents were compared. In addition, the behaviour of the model in relation to induced high intensity infiltration and to the natural infiltration was analysed. Regression analysis was performed on the long term simulation results and a linear relationship between recharge and precipitation was established.

4 SCOPE OF THE STUDY

Recharge estimates including soil moisture changes and movements have been studied by the author (Dressie, 1984) using different kinds of field experiments. Informations obtained by such field methods are limited with respect to time and space. In order to obtain a general knowledge from experimental results and to be able to extend the principles to new time periods and other areas, it is essential to have a detailed analysis of the mechanisms and patterns. This can be achieved by applying a physically based numerical model. Therefore, the scope of this study is, using a numerical simulation model:

- to simulate moisture contents, soil water flows and estimate recharge to groundwater;
- to compare simulated and measured soil moisture contents, soil water flows and estimated recharges;
- assuming that with long term simulations, recharge is more dependent on meteorological conditions than on the soil physical properties, perform statistical analysis to study the relation between seasonal recharge and precipitation.

5 NUMERICAL SIMULATION - THE SOIL MODEL

The basic equation governing the flow of moisture in the unsaturated soil profile is Richards equation and is written as follows:

$$\delta\theta/\delta t = \delta/\delta z \{K(\psi) (\delta\psi/\delta z + 1)\} + s \quad (1)$$

In equation (1):

θ = Soil moisture content by vol%

t = Time

ψ = Matric potential

K = Hydraulic conductivity

z = Soil depth

s = Source/sink term

Since, in unsaturated flow in porous media, both the conductivity and the potential are functions of the moisture content, the equation of flow is non linear. These functions vary with depth in natural profiles and boundary conditions are irregular. Consequently, no simple analytic solution exists.

The need for functional relationships between pressure and moisture content $\theta(\psi)$, and between pressure and hydraulic conductivity, $K(\psi)$, was recognized by Richards (1931). The functional relationships proposed and experimentally tested by Brooks and Corey (1964) are used in this study (equations (2) and (3)).

$$S_e = (\psi/\psi_a)^{-\lambda} \quad (2)$$

$$S_e = \frac{\theta - \theta_r}{\theta - \theta_r} \quad (3)$$

Where: S_e = Effective saturation
 θ_r = Residual soil moisture content
 ψ_a = Air entry pressure
 λ = Pore size distribution factor
 p = Porosity

and the others are as indicated in equation (1).

The most important hydraulic property governing soil water flow and the transport of solutes in a profile is the hydraulic conductivity, $K(\psi)$. There are several methods of measuring the hydraulic conductivity both in the field and in the laboratory using soil columns (Brooks and Corey, 1964; Corey et al, 1965; Laliberte et al, 1966; Hillel et al, 1970 and Malik et al, 1984). The extreme field variability makes it difficult to obtain a reliable prediction of the hydraulic conductivity of soils. Further more these methods are time consuming and very expensive (van Genuchten, 1980). Consequently, several methods of predicting hydraulic conductivities from empirical equations have been developed. Mualem (1976a) and Stephens et al (1985) have reviewed the various empirical functions used by different authors to estimate $K(\psi)$ -values. One example of such functions is Averjanovs (1950) expression of hydraulic conductivity as a power function of the effective saturation (S_e) in such a manner that:

$$K_r = K_s \cdot (S_e)^n \quad \text{where} \quad S_e = (\theta - \theta_r)/(\theta_s - \theta_r).$$

In this expression K_r is the relative conductivity, K_s and θ_s are the saturation conductivity and moisture content respectively and n is a constant. The rest of the symbols are as given earlier. This equation differs from the other empirical formulas in that it has been derived both theoretically and empirically. However, it assumes a universal n -value regardless of the soil type. Different authors have come to different values depending on how the experimental work was performed. According to Mualem (1976a), n -values of 3.5, 2.0 and 3.0 have been reported by Averjanov (1950), Yuster (1951) and Iramy (1954) respectively.

The other way of estimating the hydraulic conductivity of unsaturated soils is to use the measured capillary head - moisture content curves. This method has been used, among others, by Childs and Collins-George (1950), Millington and Quirk (1961), Burdine (1953), Brooks and Corey (1964) (Mualem, 1985). The power function in this particular case has the form of equation (2). In this equation λ is a function of the soil structure and can be estimated for a given soil type by fitting the given expression to the measured $\theta(\psi)$ curves either graphically (Brooks and Corey, 1964) or analytically by the least squares fitting method (Mualem, 1976b). According to Brooks and Corey (1964), $1.8 \leq \lambda \leq 7.3$, being small for media with a wide range of pore size and large for media with relatively uniform pore size. These semi-empirical power functions incorporate measured soil properties in the form of a pF-curve into the computation of $K(\psi)$.

In this study, estimated parameters and equation (4) are used to obtain unsaturated conductivity values.

$$K(\psi) = K_s \cdot S_e^{n + 2 + 2/\lambda} \quad (\text{Mualem (1976)}) \quad (4)$$

The simple power function on which the theory of Brooks and Corey is based, assumes a typical s-shaped $\theta(\psi)$ curve. Hence the method fails when applied to fine textured soils which usually have irregular shaped pF-curves (Mualem, 1976a). According to the same author, the method showed good results when applied to various types of soils such as silt loam, silty clay loam and clay loams and for the soil characteristic curves in the range $0 > \psi > -180$ cm water column.

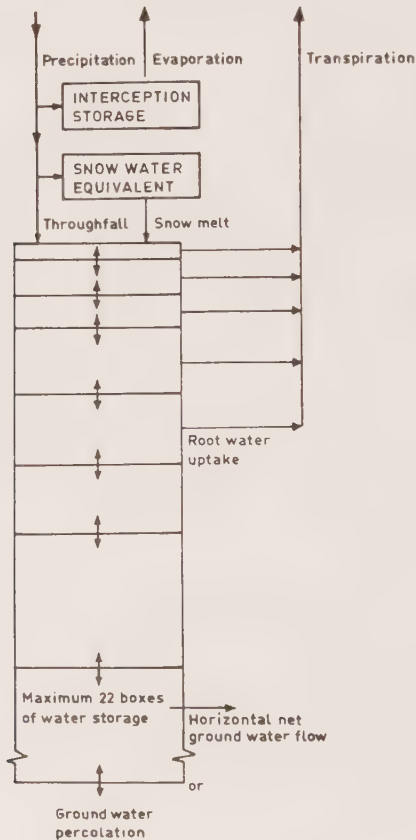
The various methods of predicting the hydraulic conductivity $K(\psi)$ from observed $\theta(\psi)$ curves assume that the pore size distribution factor λ is a unique function of the geometrical length factor of the void space. But according to Mualem (1976a, 1985), the process by which the porous media is brought to the actual moisture content (drainage or imbibition)

is also an important factor that must be considered when estimating $K(\psi)$ values. This is called the hysteretic effect. By introducing a new theory which takes account of hysteresis, Mualem (1985) found a very good agreement between theoretically and experimentally predicted hydraulic conductivities.

In the work presented here the SOIL model (Jansson and Halldin, 1980) is used to estimate recharge to groundwater. The model was developed and tried as part of the research complex in the Swedish Coniferous Forest Project (SWECON). It has been applied to different soil cover and soil types (Jansson and Halldin, (1979,1980); Jansson, (1980a, 1980b)). The flow equation is solved by finite difference approximations. The finite difference approximation requires that the soil profile is divided into discrete compartments of arbitrary thickness as demonstrated in figure 1, the mid-points of the different compartments being nodes. A linear variation of the states and the parameters that describe the hydraulic conductivity is assumed between compartments. The flow passing a certain level below the root zone is assumed to contribute to groundwater recharge.

The upper boundary at the soil surface is described by the snow melt-precipitation-interception processes and the lower boundary is expressed by the presence or absence of groundwater level. In the absence of groundwater level above the lowest boundary, gravitational force is assumed to be the sole cause of percolation at the bottom. When groundwater level is encountered above the lowest boundary of the profile, horizontal water flow is assumed.

Fig.1 The soil water flow component of the Soil model taken from Jansson and Halldin, 1980.



The potential transpiration is calculated using the Penman - Monteith equation (equation (5)).

$$L_V TR_p = \Delta((R_n - q_h^S) + \rho_a c_p \delta e / r_a) / (\Delta + \gamma(1 + r_s / r_a)) \quad (5)$$

Where: L_v = latent heat of vaporization
 TR_p = potential transpiration
 Δ = the slope of a saturated vapour pressure vs temperature curve
 R_n = net solar radiation
 q_s^h = heat flow into the soil
 ρ_a = density of air
 c_p = specific heat of air at constant pressure
 δe = vapour pressure deficit
 r_a = aerodynamic resistance
 r_s = surface resistance
 γ = psychrometer constant

The daily amount of snow melt is a function of the air temperature, solar radiation and the surface heat flow. For a detailed description on the snow dynamics and determination of the actual transpiration, the reader is referred to Jansson and Halldin (1980) pp. 13 - 15.

In general the input data in the model are classified as: (a) driving variables (b) initial values and (c) physical parameters. The driving variables are the meteorological data. In the present work the meteorological data used are temperature, precipitation, insolation, relative humidity and wind speed.

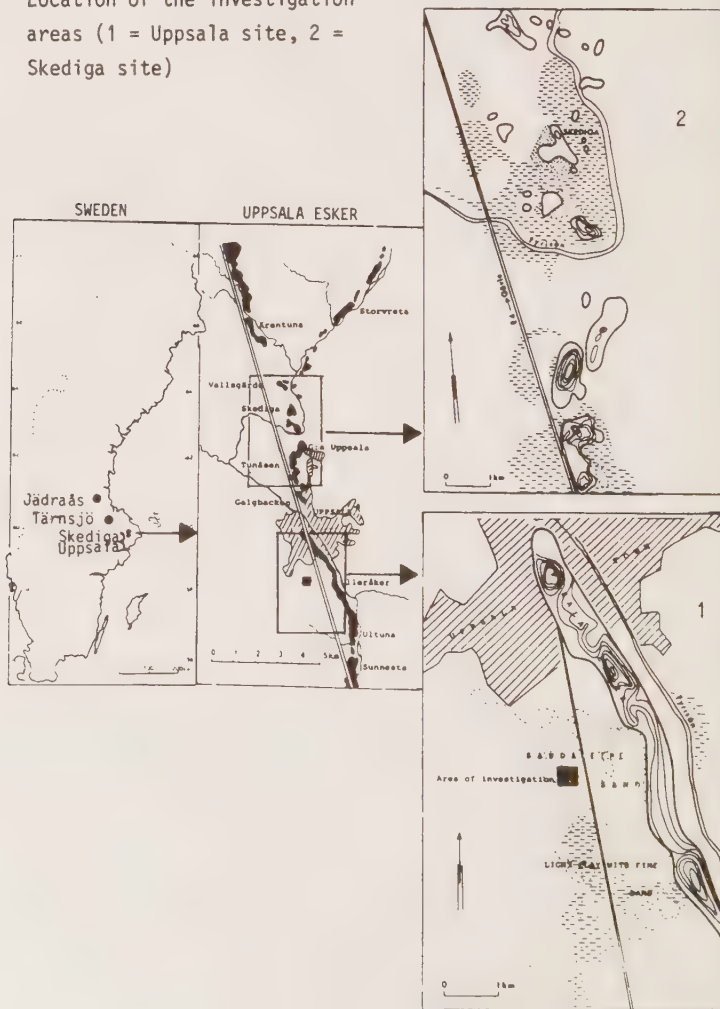
The initial soil moisture contents can either be given as a profile or can be deduced from a potential profile. Both the soil moisture content and the potential values can also be given as constants for the whole soil profile.

A complete list of all parameters and their classification is given in Jansson and Halldin (1980) pp.24.

6 AREAS OF INVESTIGATION

Simulations were made for two different sites in the vicinity of Uppsala. The simulated profiles were compared with the measured ones. The two areas of investigation are located along Uppsala Esker about 3 km south 10 km north of Uppsala respectively (fig. 2)

Fig. 2 Location of the investigation areas (1 = Uppsala site, 2 = Skediga site)

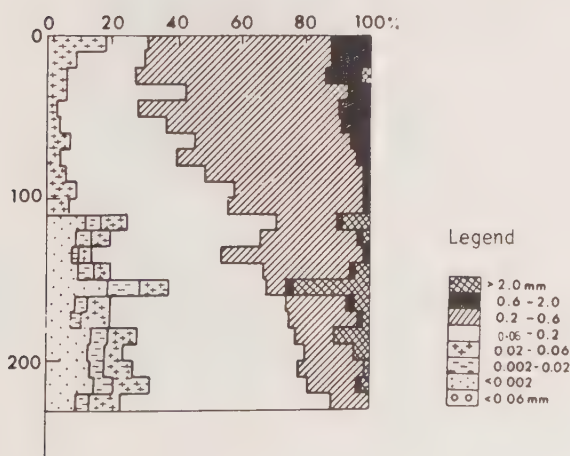


6.1 Area 1 (Uppsala site)

This area is characterized by a shallow deposit of post-glacial sand and fine sand underlain by an old Archaean formation of granite. Besides the occasional outcrops of the bedrock, it is observed that some clay lenses occur thus upsetting the homogeneity of the soil profile in the vertical direction. The top one metre layer of the soil profile is characterized by a dominance of coarser sand near the surface and an increasing dominance of finer material with depth.

The soil profile based on the particle-size distribution analysis is shown in figure 3. As can be observed from the figure, the profile at Uppsala site is dominated by sand and fine sand with 20% coarse sand at 0 - 70 cm depth. The dominance of fine sand increases with depth while that of coarse sand decreases correspondingly.

Fig. 3 The soil profile for plot A (Uppsala site)

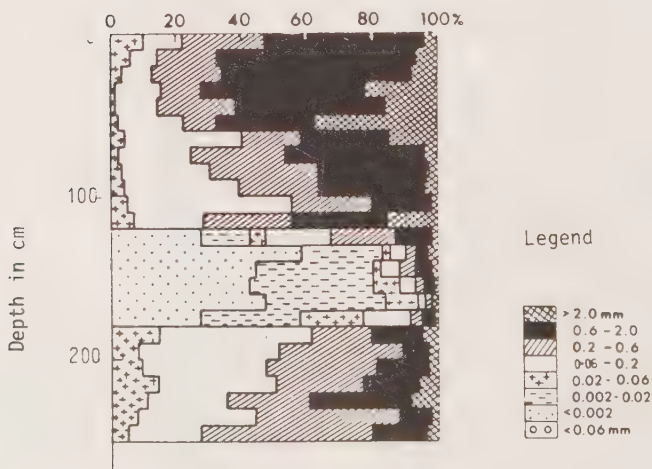


Since the sampling was done at 10 cm intervals, the thin clay lenses that occur can not be identified by the grain size analysis. However, the tritium experiment discussed later indicated a hampering layer at about 110 cm depth.

6.2 Area 2 (Skediga site)

The general stretch of Uppsala Esker is in a north-westerly direction along the fault line. There are, however, some esker formations which fall outside this general pattern and occur randomly scattered north-west of the fault line. Skediga, where plot B is located, is one of such formations. The soil profile is composed of coarse sand and gravel to 150 cm depth and a thick clay layer between 150 and 190 cm depths. This is clearly indicated in the grain size distribution analysis (fig. 4)

Fig. 4 The soil profile for plot B (Skediga site)



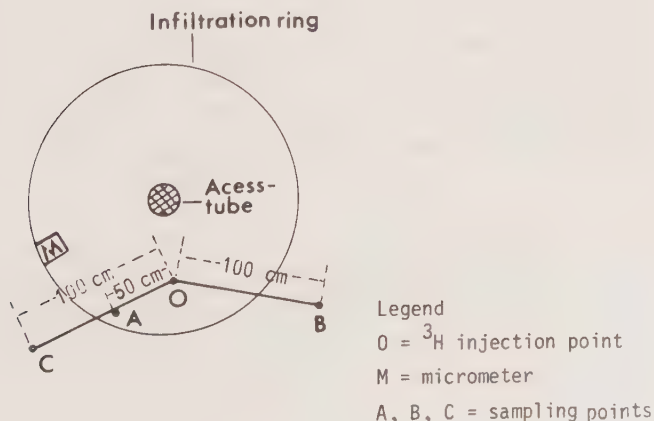
7 EXPERIMENTAL PROCEDURE

The experimental methods of soil sample coring, soil moisture monitoring and tritium profiles have been described by the author in his previous work (Dresssie, 1984). Therefore, only the infiltration scheme and the laboratory work are presented here.

7.1 Infiltration scheme

An infiltration ring, 10 cm high and 100 cm in diameter, was made out of a 2 mm thick iron plate. A PVC-tube of 12 cm height and 10 mm diameter with a notch at the upper edge to hold a micrometer was attached to the ring. The infiltration ring was then sunk to 5 cm level. The position of the ring was chosen in such a way that the access tube for neutron moisture measurement is at the center of the ring (fig. 5).

Fig. 5 Infiltration plot (Uppsala site)



The tritium injection point is marked with an (O). 60 litres of water was added in to the ring and the infiltration rate was observed. After about one hour another 60 litres of water was added and again the infiltration rate was observed. A day after another 60 litres of water was injected. The soil moisture contents were monitored at short intervals in the beginning (30 min) and with increasing intervals for about 6 days.

The rate at which the ponded water decreased in level was measured with a micrometer mounted on the upper edge of the infiltration ring. After the first 60 litres of water, the observed infiltration rate was 4.8 mm/min. When 60 more litres of water was added after about one hour, the infiltration rate observed was 5.4 mm/min.

Infiltration is a combined process of water entry through the surface of the soil, water storage and transmission through the soil. Hence anything that affects the different processes also affects the infiltration rates. Some of the factors influencing infiltration rates are the initial soil moisture contents, compaction and inwash, vegetation cover, frost heave, leaching of soluble salts, cracks due to drying, entrapped interstitial air, depth of pool at the surface and the thickness of saturated surface layer. Dry soils have a higher infiltration rate than wet soils (fig. 7). Compaction and inwash including entrapped air decrease infiltration rates while the other factors mentioned above increase infiltration rates.

It is commonly accepted that the infiltration rate decreases with time tending to reach a constant final value which is very near the saturated conductivity of the surface layer. Many equations have been proposed to describe the time dependence of the infiltration rate. The widely accepted equation is that of Philip (1957d) written as:

$$i = 1/2 St^{-1/2} + A \quad (6)$$

where i = infiltration rate

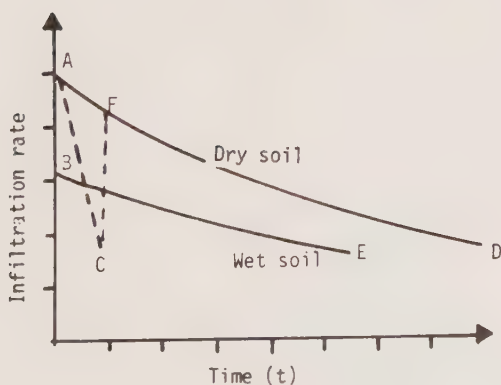
S = sorptivity

A = an estimate of i for large t values

It has also been observed (Davis de Wiest, 1967) that a secondary increase of the infiltration rate can occur due to the dissolution of the initially entrapped air in the soil (curve ACF in fig. 6). This phenomena can be the cause for the increase of the infiltration rate from 4.8 mm/min to 5.4 mm/min in this study.

Note that the discussion in section 7.1 has little bearing on the actual simulation work but is included as an additional information on infiltration processes.

Fig. 6 Schematic illustration of the infiltration process as a function of time.



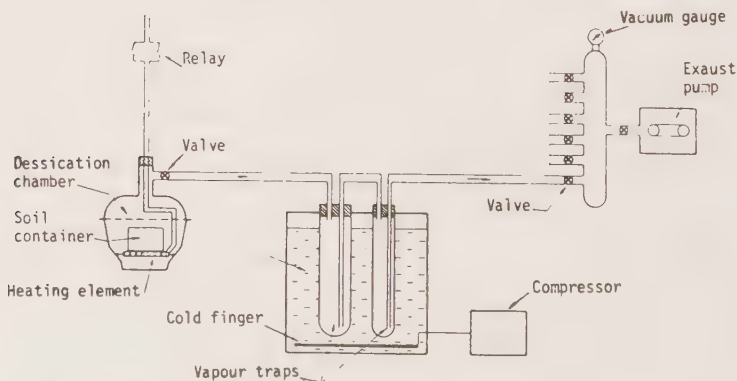
7.2 Laboratory work

The soil moisture content and the bulk density of the soil samples were determined in the laboratory by weighing the samples before and after drying in an oven at 105°C for 24 hours. The samples were then saturated and drained to obtain the moisture content - suction relationships (pF-curves). The same samples were used in estimating the saturated conductivity (K_s) for the different soil levels. This was performed by observing the volume of water passing through the saturated soil sample during an arbitrarily fixed time interval.

After the determination of the saturated conductivity, the samples were analysed for grain size distribution using sieves for coarse materials and hydrometers for finer fractions (silt and clay).

The tritium concentrations in the water samples from the injected plot were determined by the liquid scintillation spectrometry as described in the earlier work (Dressie, 1984).

Fig. 7 An apparatus for vacuum distillation of soil cores



For extracting moisture from the soil cores, the vacuum distillation method described by Saxena and Dressie (1983) was used. The set up of this method is shown in fig. 7.

8 FIELD DATA - RESULTS AND DISCUSSION

8.1 Saturated conductivity

The results of the saturated conductivity measurements for the different levels of the soil profile are presented in table 1. As observed from the table, the K_s -values lie between 13.6 and 55.5 cm/hr with an average of 32.0 cm/hr. The high value of 55.5 cm/hr at 20 to 30 cm level reflects the occurrence of a small fraction of gravel at this level while decreasing values below the 100 cm level are due to the 5 - 10% clay fraction (fig. 3).

According to Schoeller (1962), the order of magnitude of K_s can be 0.4 to 41.7 cm/hr for fine sand, 4.2 to 833.0 cm/hr for coarse sand and 4.2 to 4.1×10^3 cm/hr for gravel. Therefore, we can say that the measured conductivity values lie in the acceptable range of magnitude but whether the absolute values correspond to field conditions is questionable because laboratory experiments hardly reproduce natural conditions.

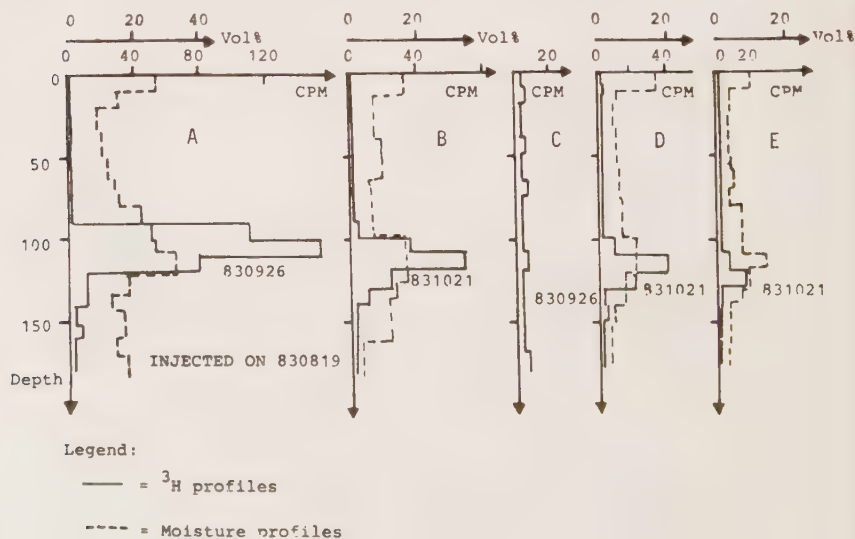
Table 1 Saturated conductivity measurements for plot A (Uppsala site).

Depth (cm)	Q(ml)	T(min)	K_s (cm/hr)	Mean
10 - 15	20	2	13.704	31.961
10 - 20	30	2	20.557	
20 - 30	81	2	55.503	
30 - 35	64	2	43.854	
30 - 40	42	2	28.779	
48 - 53	64	2	43.855	
50 - 60	73	2	50.021	
60 - 70	58	2	39.743	
80 - 90	33	2	22.612	
90 - 100	46	2	31.520	
98 - 103	35	2	23.983	
100 - 110	46	2	31.520	
120 - 130	23	2	15.760	

8.2 Tritium profiles

Injected tritium was used to trace the downward movement of soil moisture and see if the vertical flow was continuous with respect to the different layers. The tritium peak observed 38 days after injection on 830919, was found at 100 cm depth (fig. 8A). giving the rate of movement of 1.58 cm/d. Injection depth was 40 cm below the soil surface. After a further 26 days, the peak was observed to have moved 10 cm more (fig. 8B). If the rate of moisture movement were constant, the peak should have been at 141.08 cm depth. The tritium profile (fig. 8C) for a point 50 cm away from the injection point and 38 days after injection showed no trace of injected tritium indicating that the flow was vertical until then. Profiles 8D and 8E observed 64 days after injection and at points 100 cm away from the injection point indicate that during those additional 26 days lateral flow had taken place. A soil profile taken two metres away from the study plot in question showed no occurrence of clay lenses proving that such clay lenses are not extended.

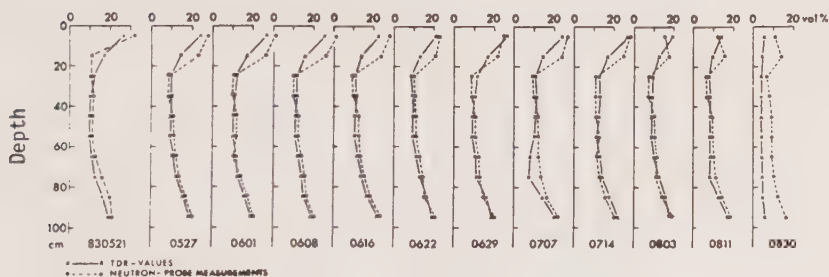
Fig. 8 Tritium and moisture profiles for plot A (Uppsala site)



8.3 Soil moisture measurements

The gravimetric method of determining soil moisture contents was used mainly for establishing the moisture content - suction relationships i.e. the pF-curves. For field monitoring of the soil moisture contents, the neutron scattering and the TDR methods were used. The results of the neutron and the TDR methods, for the periods when the measurements are simultaneously done, are shown in fig. 9 and table 2. The soil moisture contents as determined by the two methods are in good agreement except near the surface and during very dry conditions. One may question this agreement between the two methods since the moisture content did not change much with time, but Dasberg and Dalton (1984) made the same comparison for extremely variable conditions and showed an almost 1:1 relationship between moisture contents measured by the TDR and the neutron methods.

Fig. 9 Soil moisture contents as measured by the neutron probe and the TDR methods for plot A (Uppsala site).



The deviations near the surface are due to the failure of the calibration curve to fit the surface layer. In this work the same calibration curve was used for the whole profile. In order to get better results with the neutron method, each soil layer should have its own calibration curve. It has been observed that calibration curves to surface layers of the soil are difficult to obtain because of the incomplete reflection of the fast neutrons (Schudel, 1982).

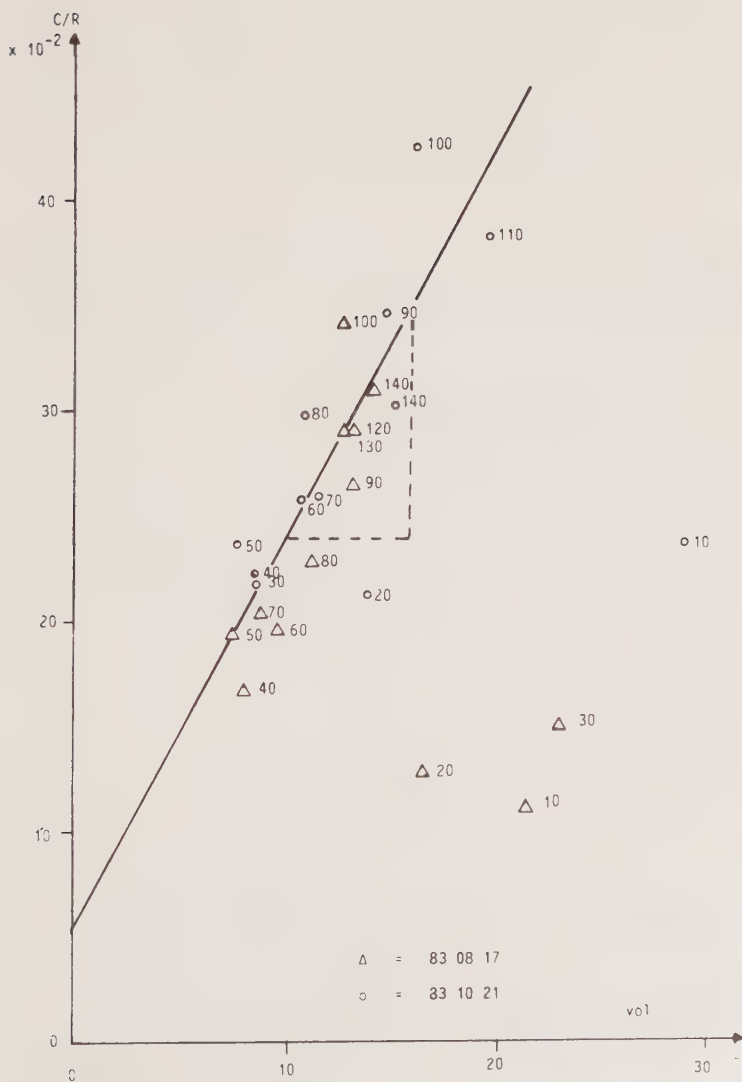
Concerning the TDR method, the presence of air gaps between the soil and the electrodes can decrease the dielectric conductivity thus affecting the moisture content (Topp et al, 1982; Dasberg et al, 1984). The dielectric conductivities of pure air, soil and water are 1, 4 and 80 respectively (Dasberg et al, 1984). This can be the reason why the TDR method underestimates the moisture contents during the dry periods when air fills the gap between the electrodes and the soil. These gaps are caused by the freezing and thawing of the soil moisture during winter.

Fig. 10 shows the calibration curve used for the whole profile, plot A. In the figure C stands for the actual counting rate observed at the different levels and times while R is the reference count rate measured with the electron source in the protecting shield. The numbers besides the points stand for the depths below the soil surface where soil cores are taken. As seen from the figure, the scattered points are those for the surface layer, showing the problem in estimating moisture content near the surface of the soil profile.

Table 2 Soil moisture measurements by the neutron scattering and the TDR methods for plot A (Uppsala)

DATE	METHOD	0-10	10-20	20-30	30-40	40-50	50-60	60-70	70-80	80-90	90-100 cm
1903	TDR	26.4	16.9	11.7	10.7	10.2	9.9	11.2	12.8	16.1	19.4
0521	NEUTRON	31.5	10.6	10.4	10.6	11.4	11.4	12.9	15.3	19.9	19.7
	TDR	24.0	14.2	9.8	9.8	9.6	9.5	11.0	12.5	15.6	18.6
0527	NEUTRON	27.5	23.0	9.2	9.7	10.5	11.1	11.3	12.7	15.1	20.0
	TDR	27.0	17.0	11.8	11.0	10.5	10.1	11.2	12.6	15.8	19.0
0601	NEUTRON	31.1	26.1	10.7	10.7	11.1	11.3	11.5	12.8	15.1	19.3
	TDR	25.4	15.7	11.7	11.7	11.4	11.0	12.5	14.4	18.1	21.8
0608	NEUTRON	28.8	25.3	10.9	11.3	12.1	12.6	12.9	14.7	17.6	22.3
	TDR	22.5	13.6	10.2	10.6	10.4	10.3	12.0	13.7	17.6	21.4
0616	NEUTRON	27.3	23.4	9.9	10.4	11.6	12.1	12.4	14.4	17.6	22.3
	TDR	20.9	13.1	8.6	9.5	9.9	9.8	11.4	13.1	16.6	20.0
0622	NEUTRON	21.3	20.3	8.8	9.5	10.6	11.3	12.0	13.6	16.0	20.9
	TDR	26.6	16.9	11.4	10.3	9.8	9.5	10.7	12.2	15.6	18.9
0629	NEUTRON	24.8	21.7	8.5	9.2	10.1	10.6	11.3	12.9	15.4	20.0
	TDR	23.9	14.8	10.9	11.0	10.8	10.4	8.8	7.7	14.2	20.6
0707	NEUTRON	26.5	23.2	9.8	10.2	11.2	11.9	11.9	13.5	16.0	21.2
	TDR	26.6	17.1	12.8	12.6	12.1	11.6	12.7	14.2	18.0	21.8
0714	NEUTRON	27.6	24.8	10.6	10.7	11.3	12.0	12.0	13.8	15.9	21.1
	TDR	19.3	12.6	9.0	9.6	9.7	9.6	10.9	12.4	15.4	18.3
0803	NEUTRON	15.2	17.5	7.7	8.5	9.8	10.4	10.4	12.2	14.6	19.4
	TDR	14.0	9.4	7.6	8.4	8.5	8.3	8.4	8.6	12.9	17.1
0811	NEUTRON	12.3	15.5	7.0	8.0	9.7	9.9	10.5	11.5	14.0	18.8
	TDR	5.4	4.6	4.3	4.4	4.4	4.4	4.7	4.9	5.3	5.6
0830	NEUTRON	10.9	13.8	6.5	7.8	9.1	9.7	9.7	10.7	12.5	16.5

Fig. 10 A calibration curve for the neutron moisture probe and for plot A (Uppsala site)



9 INFILTRATION PROCESS SIMULATION

Previously done tracer experiments (Dressie, 1984) have indicated that when moisture movement in the upper soil profile was studied, intensive infiltration was sometimes not identified proving the existence of two domains of moisture movement (Germann and Beven, 1981), namely, flow in micropores (capillary pores) and flow in macropores (non-capillary pores). Flow in macropores is extremely fast at or near saturation. During the tracer experiment, when the front in the micropores is strictly followed, flow in the macropores may be hard to detect.

If the penetration rate of the moisture front in soil profile were the same at all points within a unit area, specially at high degree of saturations, then the observed and simulated profiles should look alike. On the other hand, if preferentially faster flows occur in bigger pores, then the observed and simulated profiles will be different as regards the depth of front penetration since mathematical models are based on the one dimensional solution of Darcy's equation which in turn requires laminar flow. In order to look into this problem more closely, a high intensity infiltration was simulated using the mathematical model discussed in section 5. In the second part of the simulation work, the long term soil moisture flows with the natural precipitation were simulated and used to estimate the ground water recharge. Before the layout of the simulation is discussed, it is appropriate to review some views on the effects of macropores on soil water flow.

9.1 Macropores and soil water flow

Many soil water models are based on the Richards equation whose solution requires laminar flow in the smaller (capillary) soil pores. The presence and importance of macropores in soil water flow has been of prime interest for many workers and an extensive work in this field has been preceeding very fast during the last few years. A summary of this work was done by Beven and Germann (1982). Although, theoretical and experimental studies (Germann and Beven, 1981(I,II), Beven and Germann, 1981(II), 1982, Davidson, 1985, Wang and Narasimhan, 1985) confirm that the penetration of soil water

flow is enhanced by non-capillary pores, the identification and effect of of macropores on soil water flow has had very little attention in most simulation works.

Beven and Germann (1982) discussed the morphological groupings of macropores which included pores formed by plant roots, pores formed by soil fauna, cracks and fissures and natural soil pipes.

Big pores can be formed by dead or living plant roots and are tubular in shape. In forest soils such pores can contribute up to 35% of the total soil volume.

Macropores formed by soil fauna such as earthworms and ants are also tubular and their size can range from less than 1 mm to over 50 mm and can extend to a depth of 100 cm. Earthworms dominate in low acid to neutral soils and insects like ants prefer acid soils.

Seasonal drying (desiccation) and wetting of clayey soils, chemical weathering, thawing and freezing all cause soil cracks and fissures.

Subsurface erosive action of soil water flow in highly permeable and non-cohesive soil materials can cause pipings.

Although, macropores contribute a minor part of the total porosity, they have a very big influence on the saturated hydraulic conductivity of soils (Beven and Germann, 1982). Macropores are specially effective at very high degree of saturation (positive) and at very dry conditions (negative). Hence, macropores enhance flow at saturation and impede flow at very low saturation.

Using a combined flow model, Beven and Germann (1981) showed that:

- (a) In soils of low hydraulic conductivity the presence of macropores increases the depth of infiltration while in the absence of macropores the infiltration depth is extremely limited (fig. 11A).
- (b) In soils of very high hydraulic conductivity, the effect of the presence or absence of macropores is minimal (fig. 11B).

Fig. 11A The effect of the presence (a) and absence (b) of macropores on infiltration in soils of relatively low hydraulic conductivity (After Beven and Germann, 1981).

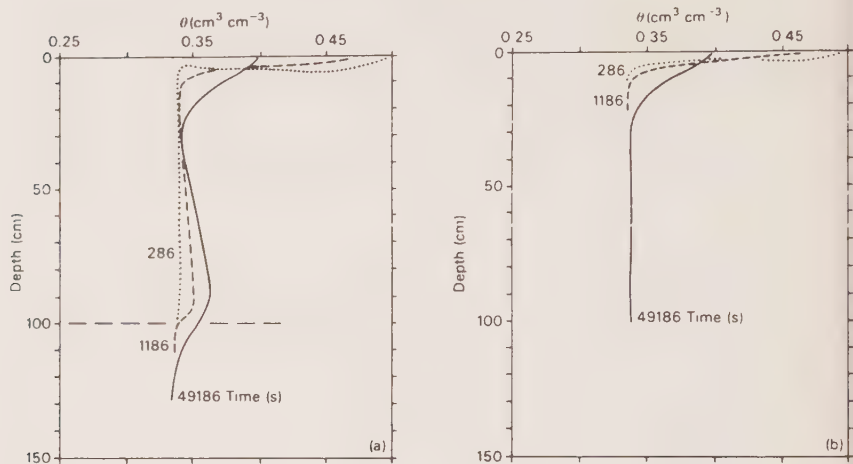
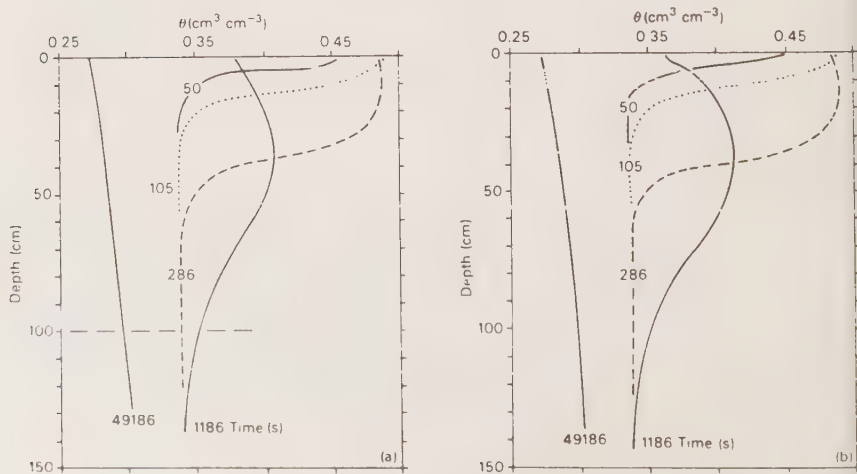


Fig. 11B The effect of the absence (a) and presence (b) of macropores on infiltration in soils of relatively high hydraulic conductivity. (After Beven and Germann, 1981)



The concept of "domain" in soil water flow was introduced by Beven and Germann (1981) according to which two types of flow are distinguished, namely, flow in micropores as described by Richards and Darcy's equation and bulk flow in macropores

The boundary between macro- and micropores is a matter of opinion and can be estimated from the equation that relates a cylindrical pore radius (R) to the potential (ψ) (Germann and Beven, 1981) as shown in the equation below.

$$\psi = -\frac{2\sigma}{R\rho_w g} \quad \text{where } \sigma = \text{surface tension}$$

$\rho = \text{density of water}$

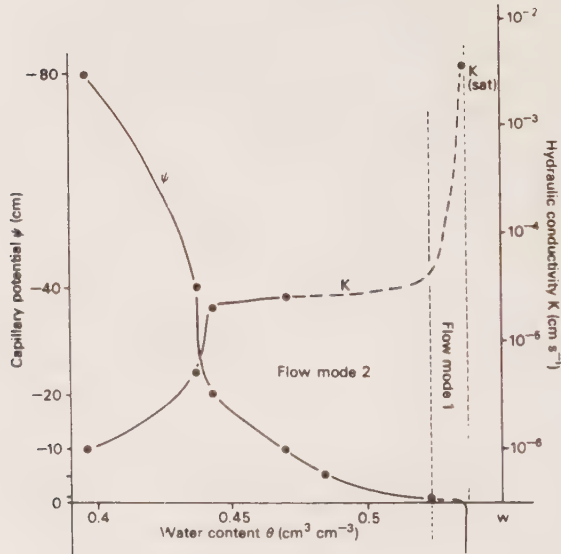
$R = \text{pore radius}$

$\psi = \text{potential in units of length}$

$g = \text{gravitational acceleration.}$

The continuity of big pores is usually disturbed by constrictions (pore necks) which in turn makes the definition of big pores in terms of pore geometry difficult because of the complexity of pore sizes. That is why an indirect way of defining macropores using the relation of moisture retention curve to pore sizes, as shown above, is used. Such an indirect of defining pore sizes is rather arbitrary and different values have been proposed by different authors (Beven and Germann, 1982). In the equation shown, the authors used an arbitrary value of -1.0 cm for ψ and the corresponding pore radius of 0.15 cm as a boundary between the two flow domains. Typical combined flow mode illustrating the two domains is shown in fig. 11C. The authors used data from Brühlhart (1969) to construct flow mode 2. The figure shows that at high saturation soil water flow is independent of the potential and because of the fast emptying of the big pores, the hydraulic conductivity falls steeply (mode 1). At higher potentials, the hydraulic conductivity is constant until the air phase in the macropores is continuous after which the hydraulic conductivity starts to fall with falling soil water potential (mode 2).

Fig. 11C Schematic presentation of flow domains. Adopted from Germann and Beven (1981)



9.2 Parameters

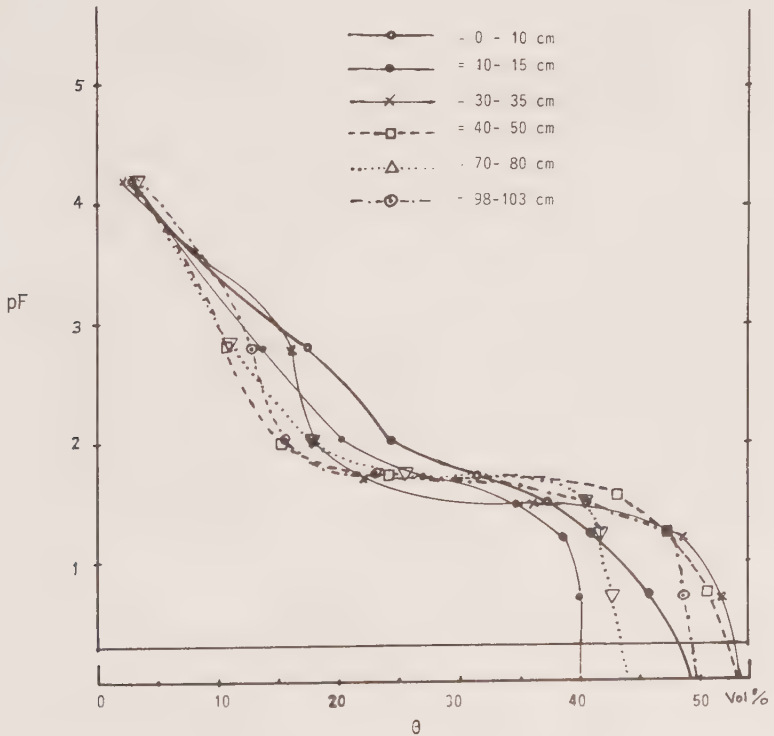
9.2.1 Plot A (Uppsala site)

9.2.1.1 The soil characteristic curves (pF-curves)

The pF-curves for the different levels and for Uppsala site were evaluated in the laboratory and presented in fig. 12A. As observed in the figure the curves deviate from each other at soil moisture contents below 25% and above 35% by volume. These pF-curves are used for the estimation of the necessary soil physical properties such as the air entry pressure (ψ_a),

the residual moisture content (θ_r), the pore size distribution factor (λ) and the porosity (p) in equations (2) and (3). The parameter θ_r represents the soil moisture content at $d\psi/d\theta = 0$ (see fig. 12B). For practical purposes θ_r was assumed to be the moisture content at the permanent wilting point of 15 000 cm water column (van Genuchten, 1980).

Fig. 12A pF-curves for the different levels of plot A (Uppsala site)



The parameters ψ_a and λ can be determined from the graphical representation of effective saturation S_e versus suction ψ . In the graph (fig. 13), ψ_a is the suction when $S_e = 1$, and λ is the slope of the curve. The tortuosity factor $n (= L_e/L)$ varies between 0 at $\theta = \theta_r$ and 1 at $\theta = \theta_s$ (saturation). This was observed by Burdine (1953) and Willie & Gardner (1958)

(Brooks and Corey, 1964). The same authors refer to Carman (1937) who found that the tortuosity of granular and isotropic soils when fully saturated, is equal to 2.0. Mualem (1976) found that n was dependent on soil texture but he considered it as independent of the soil water content. In the expression for n , L is the total length travelled by the water particle and L_e is the effective length (shortest distance between two points).

Fig. 12B A sketch of a pF-curve showing θ_r and θ_s values.

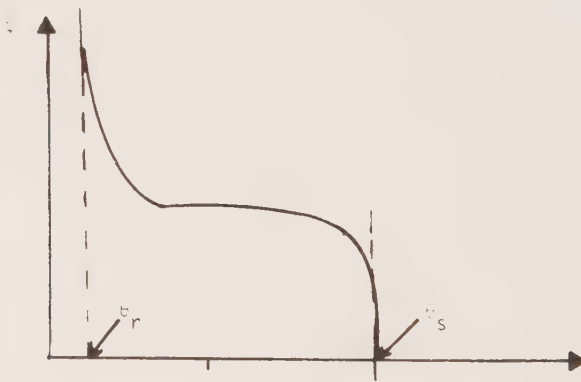
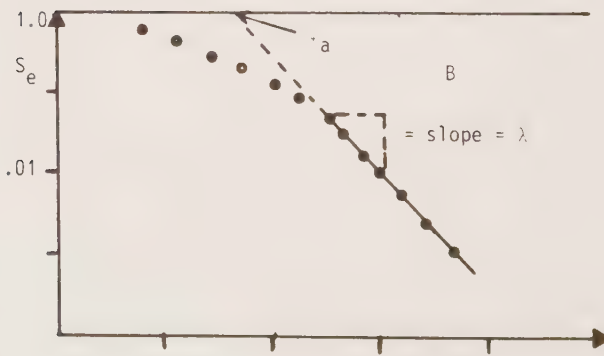


Fig. 13 Effective saturation (S_e) as a function of suction (ψ)



Using the estimated parameters and equation (4), the unsaturated conductivities were calculated. The calculated soil characteristic and tension curves are shown in fig 14. Fig. 15 shows the estimated and observed pF-curves. As seen from the figure the agreement between observed and estimated values is quite good except for the region near saturation. This discrepancy is in order because the method is valid for the suction range $> \psi_a$ (Brooks and Corey, 1964).

Because of the effect of flow in macropores and the resulting fingering phenomena, the measured pF-curves had to be modified so that the observed soil moisture profiles (using the neutron meter) are in agreement with the simulated ones. These modified pF-curves and the corresponding hydraulic conductivities are shown in fig. 16. For the long time soil water flow simulations, however, the measured, unmodified pF-curves were used.

9.2.1.2 Meteorological data

The driving variables in the model, as stated earlier, are the meteorological data namely precipitation, air temperature, insolation, relative humidity and wind speed. The evapotranspiration values were evaluated by using this data in the Penman formula. For Uppsala site data from Ultuna, 2 to 3 km from the study site was used. The precipitation (monthly sums) and air temperature values for the study period are presented in figs. 17A and B. One can observe that 1982 and 1984 were relatively drier years than 1980, 1981 and 1983.

Fig. 14 Estimated pF-curves (A) and hydraulic conductivity versus tension curves (B) for the soil profile at plot A (Uppsala site)

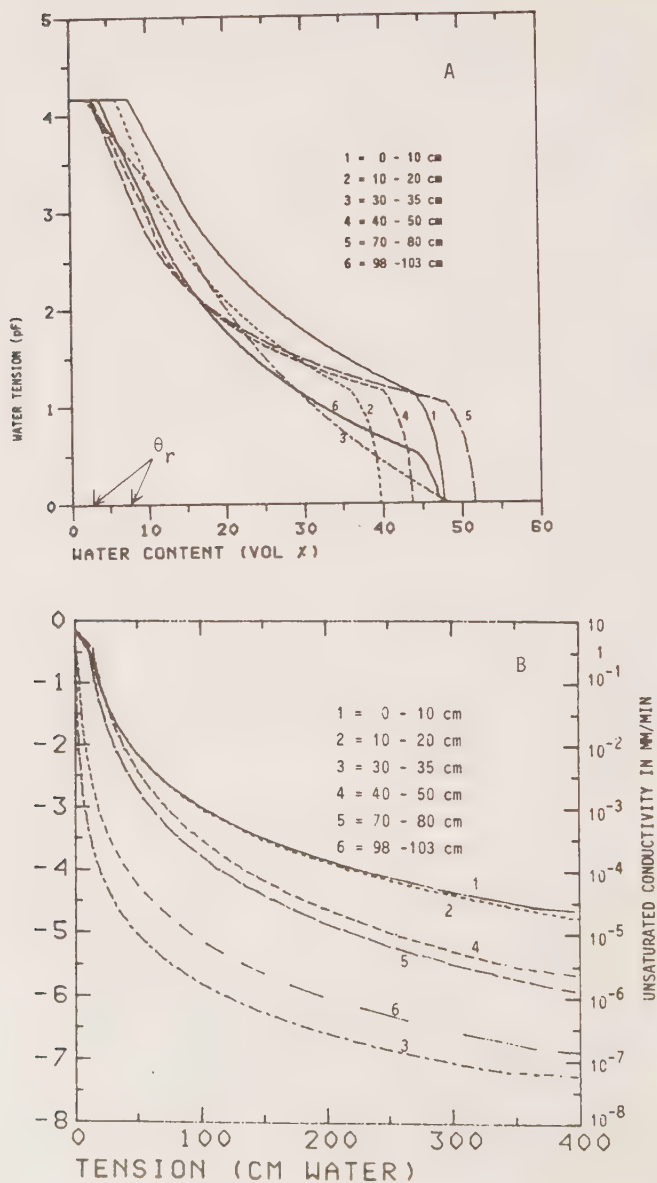
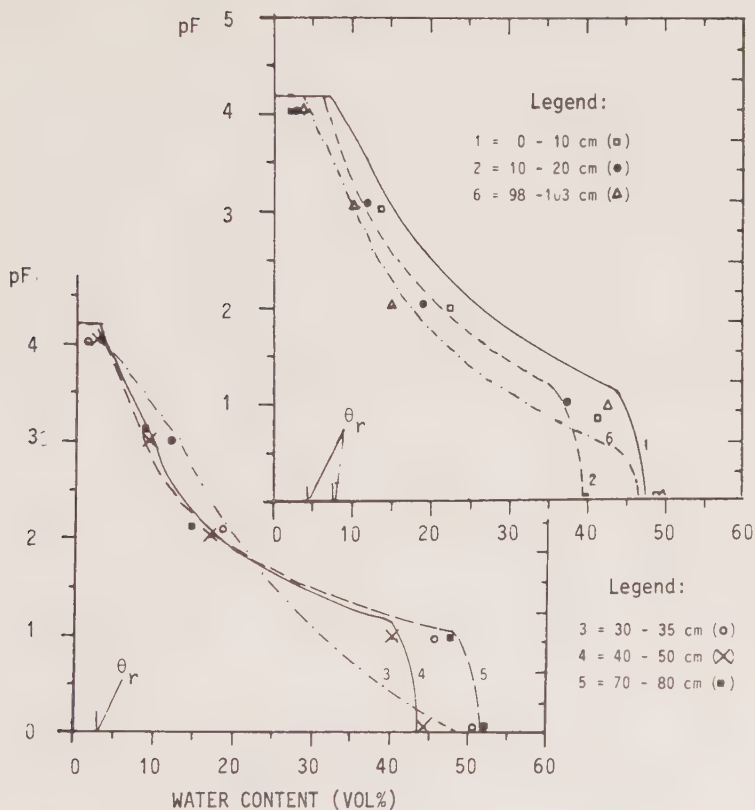


Fig. 15 Estimated pF-curves and measured points for plot A (Uppsala site)



Using the artificially added infiltration water as an input, and neglecting the evapotranspiration loss, the soil water contents in the profile were simulated. The simulated and observed soil moisture profiles are shown in figs. 16A - D.

Fig. 1 δ The modified pF-curves (A) and the tension versus conductivity Curves (B) for plot A (Uppsala site)

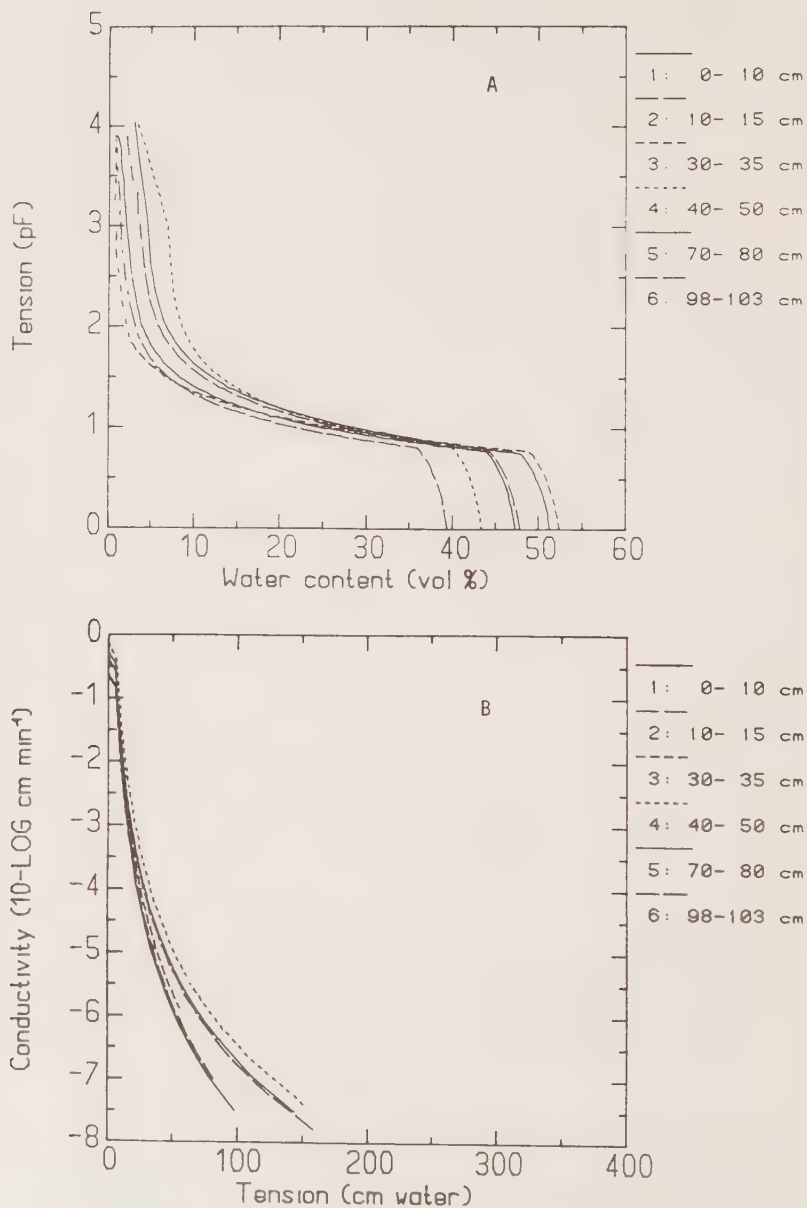
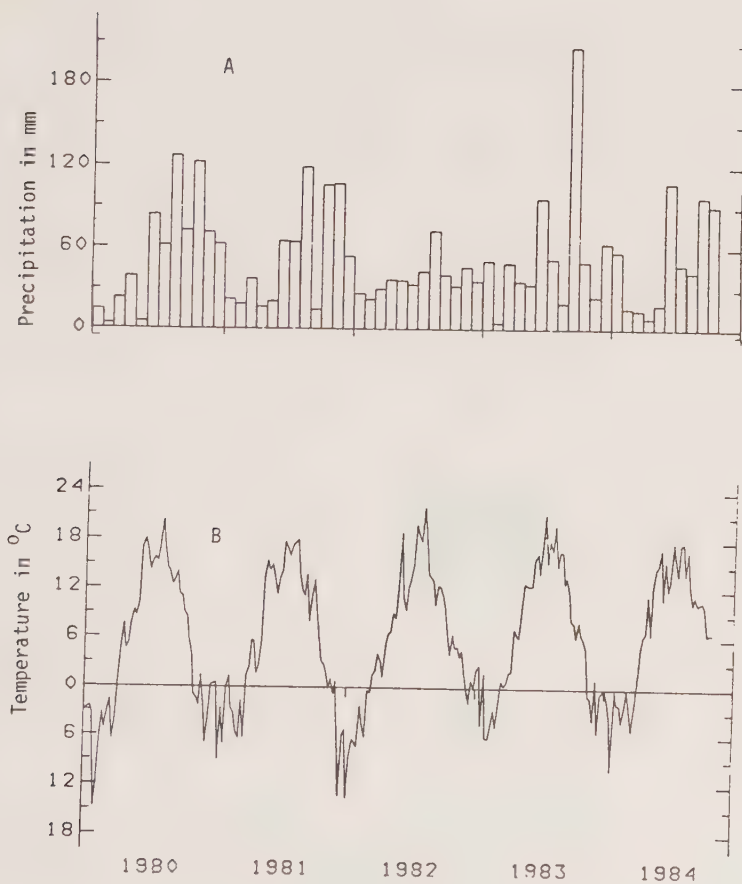


Fig. 17 Monthly precipitation (A) and 5-day mean temperatures (B) for Uppsala site.

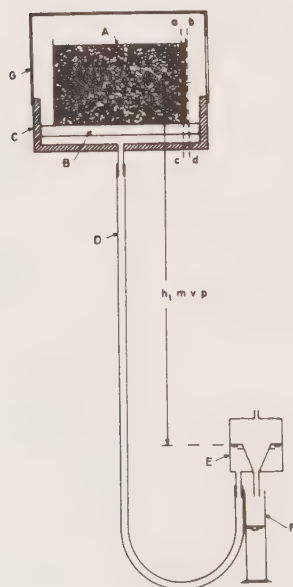


9.2.2 Plot B (Skediga site)

9.2.2.1 The soil characteristic curves (pF-curves)

The most widely used technique for the laboratory evaluation of the soil moisture characteristic curves (pF-curves) is the pressure plate and pressure membrane water extraction method. The simplest type of a pressure plate apparatus is shown in figure 18A. In this method, the soil sample (A) is placed on a porous plate (B) whose pores form a continuous system with the soil pores. The soil sample with the porous plate are enclosed in plexi-glass cover (C, G). The lower side of the porous plate is provided with an outlet which connects it to free water at atmospheric pressure (E).

Fig 18A Pressure plate apparatus for the evaluation of the soil characteristic curves (After Andersson and Wiklert 1972)



The pressure difference across the porous plate is controlled by the hanging water column (0-1 bar) and by vacuum (higher suction range). The extracted water is received in the graduated cylinder trap (F). The design of the pressure chamber can vary depending on the suction range to be evaluated. At this point, it is appropriate to mention that the soil moisture characteristic curve is hysteretic and the value of suction at certain given moisture content is not unique being higher during desorption and lower during sorption, thus forming a "hysteretic-loop", the causes of which are the inkbottle, the contact angle, the entrapped air and

the swelling and shrinking effects (Hillel, 1971). The pF curves for plot A were evaluated in the lab using the above method. Such data lacks for

Skediga site and had to be determined indirectly using the particle size distribution data. Moisture characteristic curves for a broad spectrum of Swedish soils, ranging from sand to heavy clays (85% clay content) and representing many different geographical regions, have been evaluated by Andersson and Wiklert (1972). The soils are classified into topsoils (0 - 10 cm) and subsoils (10 - 100 cm) with further subdivision according to clay content by weight percent. By matching the grain size distributions of the samples from the study site with those presented by the above authors, the corresponding pF-curves were selected as shown in fig. 18B. These pF-curves are used to estimate the necessary parameters. The estimated and simulated values are compared (fig. 18B) and corresponding hydraulic conductivities are presented in fig. 18C.

9.2.2.2 Meteorological data

For Skediga site, meteorological data from F16 (the air force weather station very near the study area) was used. The monthly sums of precipitation and the 5-day mean temperatures for the study period (1960 - 1980) are shown in figs. 18D and E. The data after 1980 was not included in the figures because of the lack of space only.

9.3 High intensity infiltration process - Results and discussion

The observed (fig. 19A) and simulated (fig. 19B) profiles after the first infiltration indicate that the soil moisture front after 40 hours and 15 minutes was at the 130 and 50 cm depths respectively. This gives a percolation rates of 3.23 cm/h and 1.24 cm/h for observed and simulated profiles respectively. The observed (fig. 19C) and simulated (fig. 19D) profiles after the second infiltration show that in both cases, the front had penetrated to the 130 cm depth after 7 hours and 15 minutes giving a percolation rate of 17.93 cm/h. This very high percolation rate indicates that during the second infiltration, the soil was merely draining without any redistribution taking place.

Fig. 18B and C pF-curves (A) and unsaturated conductivities (B) for Skediga plot. The marked points on the pF-curves are measured values. The different curves in the figures below correspond to the different soil layers as shown in the legend.

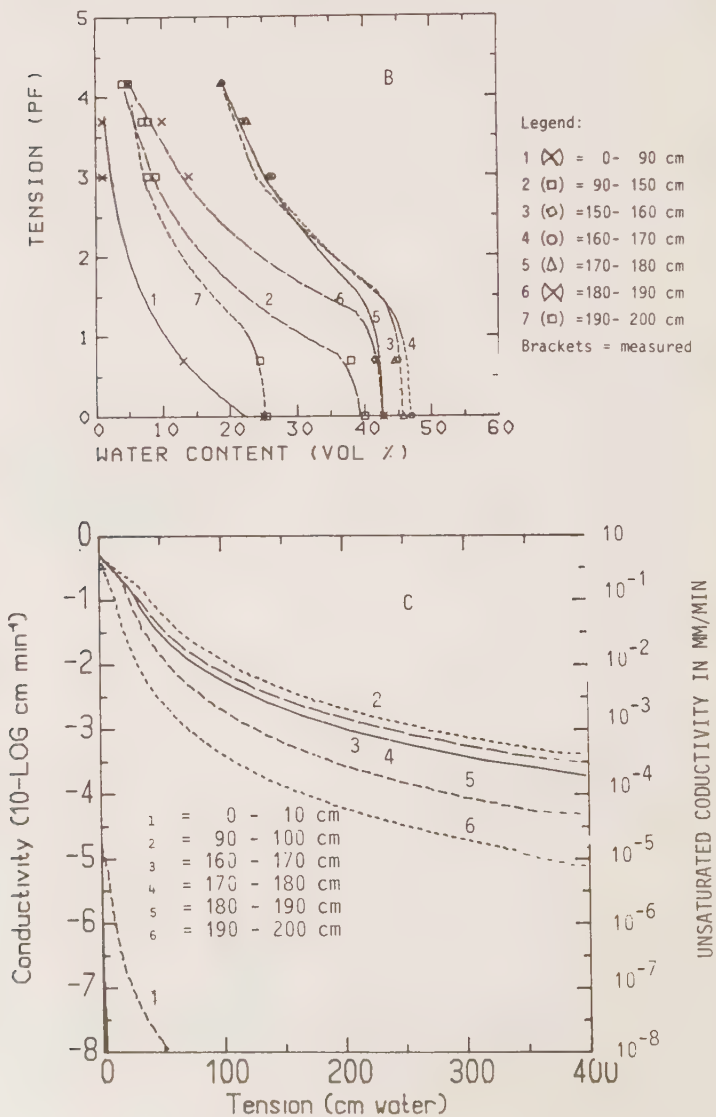


Fig. 18D Monthly precipitation (A) and 5-day mean temperatures (B) for Skediga site, 1961 - 1970.

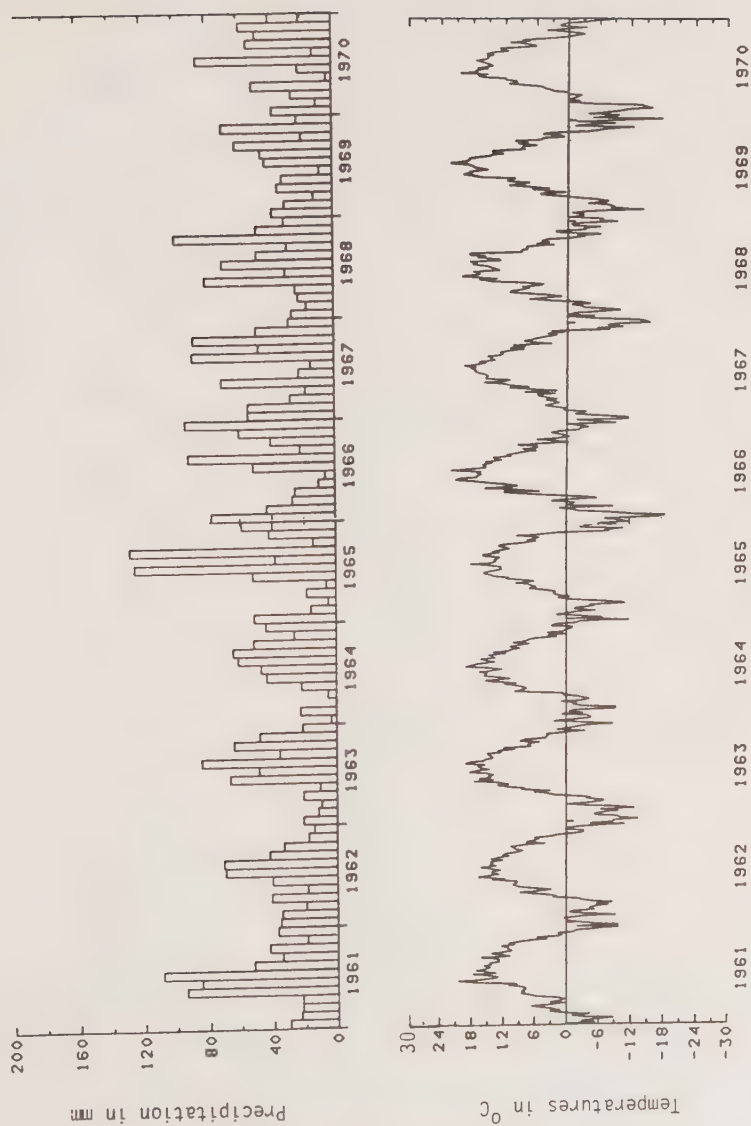


Fig 18E Monthly precipitation (A) and 5-day mean temperatures (B) for Skwdiga site. 1971 - 1980.

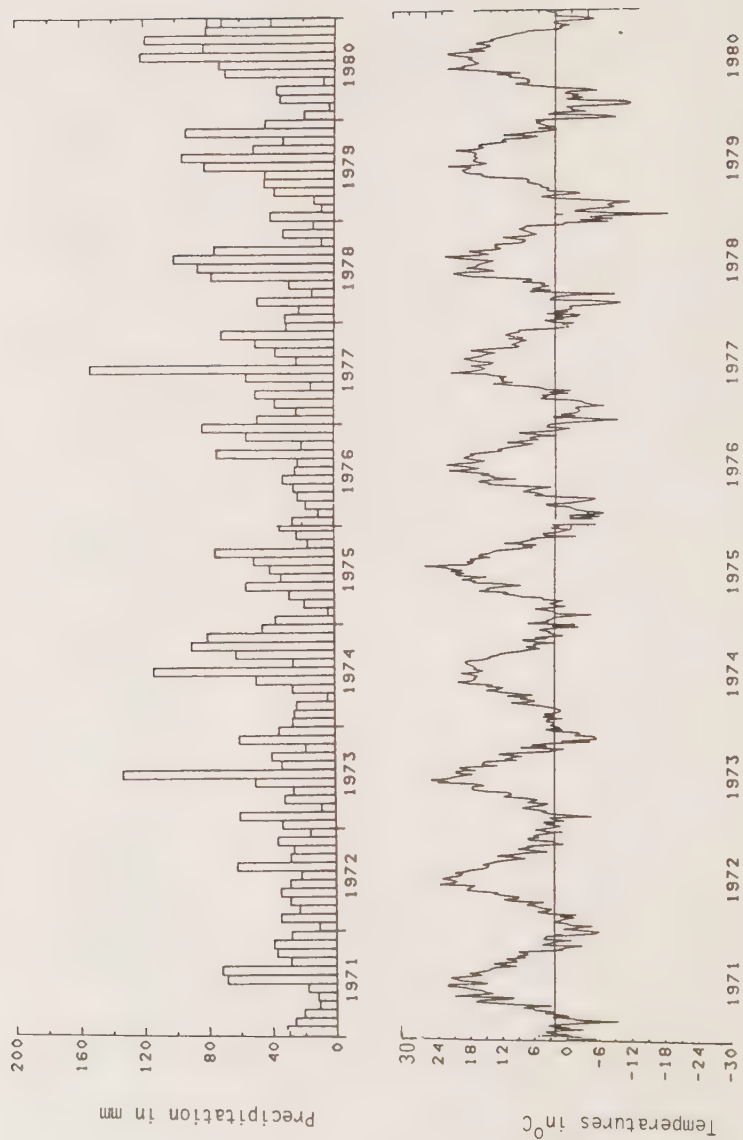
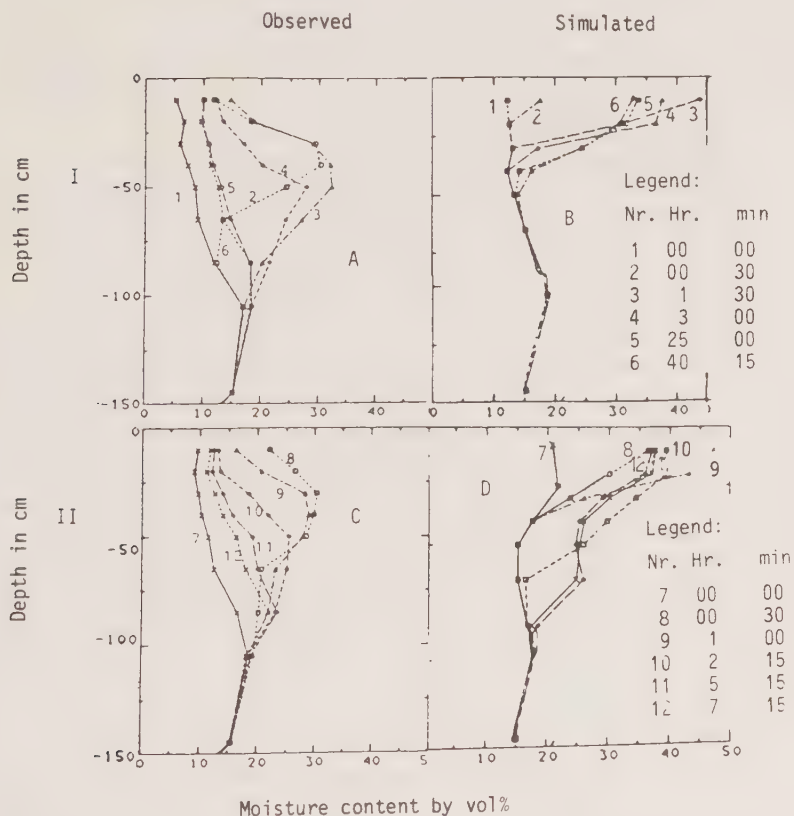


Fig. 19A - D Observed and simulated moisture profiles after the first (I) and the second (II) infiltration for plot A (Uppsala site)



The initial soil moisture profile was defined by a variable potential profile and as can be observed from the figures, the initial soil moisture profiles are similar for the measured and simulated cases prior to the first infiltration event but quite different prior to the second infiltration event. The comparison of the simulated and observed profiles showed quite different patterns of flow conditions. In other words, the similarity between the observed and simulated profiles was poor indicating that the soil physical properties and the boundary conditions used in the simulation model did not completely describe the natural infiltration/percolation processes in the field.

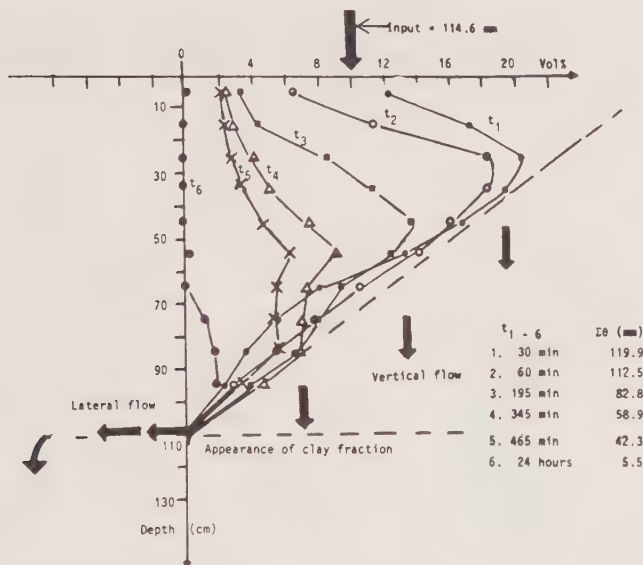
In natural conditions the wetting front does not move in the same way at all points because water flows preferentially via vertical inhomogeneities such as cracks, root channels and small highly permeable areas. This causes rapid water movement at some points and relatively slow movement at others. The phenomena is called "fingering" and is especially important when structural and textural changes in the soil profile cause instability whereby the water-course takes different ways. A typical example is when water passes from fine textured horizon into a coarser layer. This fingering effect, which is common under field conditions is not accounted for in the model and is indicated by the deeper wetting front in the observed profile as compared to the simulated one. In the simulated case the moisture is maintained near the surface. The main cause of which is the differences in water holding properties in the top soil. As observed, the measurements indicated a coarse sand layer near the surface whereas the simulations reflect a substantially finer material. It is important to mention that, when comparing simulated and measured profiles, one should note that there are some fundamental draw backs with the observation methods. For instance, considering the neutron scattering method, besides the problem of constructing correct calibration curves, the method fails to give accurate results near the surface of the soil profile due to the escape of the fast neutrons to the atmosphere. For example, curve number 6 in fig. 19A which represents a profile drained only by gravitational force after the infiltration should correspond to tensions below or equal to 2 on the pF-scale. According to the pF-curve in fig. 12A, this corresponds to moisture contents above 20% which is substantially more than the neutron probe measurements.

One important characteristic of glaciofluvial deposits is the occurrence of clay lenses in the profile. Such clay lenses are usually very thin and have a limited lateral extent so that soil water impeded by them can finally find its way down to deeper layers.

The infiltration/percolation process after the second infiltration event

was further analysed using fig. 19E in which the differences between the initial and the total soil moisture contents at times t_1 to t_6 after the infiltration are drawn along the profile.

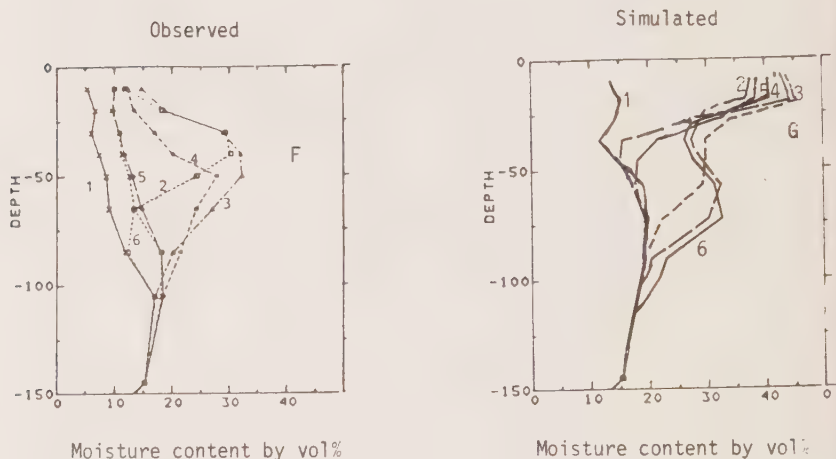
Fig 19E The movement of infiltrated water in the soil profile for plot A (Uppsala site).



The infiltrated water (114.6 mm) and the maximum soil moisture content observed in the soil profile after 30 minutes (119.9 mm) compare very well. The deviation is only 5%. There occurs no increase in the soil moisture content below the 110 cm level. This is the level where the clay layer appears. At the same time, after 24 hours, almost all the added moisture had left the profile. Then there is no doubt that at about the 110 cm level, lateral flow takes place which, at some distance from the observation plot may change to vertical flow since such clay lenses are not extensive laterally.

In order to account for and estimate the effect of piping, the infiltration rate was increased threefold (I_3) so that the effective infiltration $I_{\text{eff}} = 1/3 I_3 + 2/3 I_0$ where I_0 is the initial infiltration. The threefold increase was decided upon by trial and error until the observed and simulated soil moisture profiles coincided (figs. 19F - G).

Figs. 19F and G Observed (first infiltration) and simulated (increased infiltration) moisture profiles for plot A (Uppsala site)

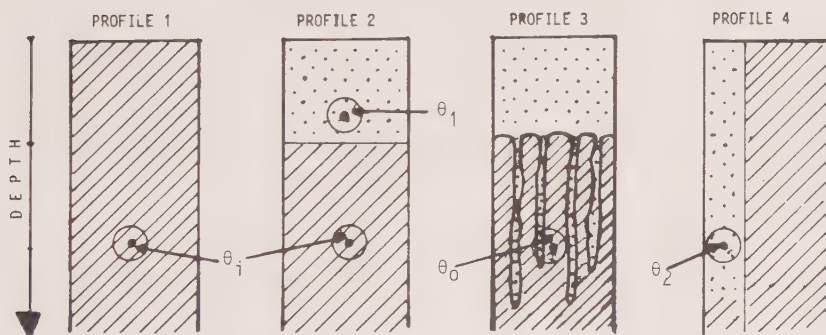


Note that the pattern of moisture movement is still different. This phenomena can be discussed in the following manner. Assuming that the pF-curves constructed using the measured saturated conductivities and the neutron moisture measurements are correct, a soil moisture flow pattern as exhibited in fig. 20 can be theoretically postulated. In the

figure, profile 1 represents conditions before infiltration took place. The initial moisture content is denoted by θ_i . The simulated soil moisture profile with its shallow depth of infiltrated moisture front is presented in profile 2. The moisture content for this case is θ_1 . A sketch showing the observed field conditions with fingering effects is depicted in profile 3.

The observed soil moisture content has the symbol θ_o and near the moisture front, it reflects both the initial moisture content and the effect of fingering. Increasing the intensity of input threefold is the same as having

Fig. 20. Hypothetical presentation of moisture flow with fingering effects.



the original input on 1/3 of the surface of the study plot. This is shown in profile 4. The simulated moisture content after threefold increased intensity is denoted by θ_2 . When the whole profile is considered, then, the soil moisture content contributed by infiltration is only 1/3 of θ_2 . Therefore one can say that the observed moisture content is the sum of one third of infiltrated and two thirds of the initial moisture content and can be expressed as:

$$\theta_0 = 1/3 \theta_2 + 2/3 \theta_i \quad (7)$$

As mentioned before, this discussion is based on the assumption that the saturated conductivities determined in the laboratory correctly reflect field conditions and that the neutron soil moisture measurements were accurate. But if, on the other hand, we assume that the measured saturated conductivities and consequently the pF-curves do not describe the field plot completely, which is a legitimate assumption since laboratory methods usually give higher values as a result of the saturation procedure and the measuring of the desorption cycle only, then the discussion can be carried out in a different manner as follows.

Since the model responds to the soil properties that have been defined by the user, the dissimilarities between the observed and simulated profiles can either be due to errors introduced when measuring pF-curves or due to errors in the neutron soil moisture measurements. Since the latter can not be proved, and since it was indicated that there was a discrepancy between the pF-curves for the field plot and those used in the simulations, it became clear that the problem can best be tackled through modification of the pF-curves used in the simulation.

Assuming that the neutron measurements are correct, new pF-curves for a better description of the soil physical properties were constructed from the soil moisture contents. These pF-curves and the corresponding tension versus hydraulic conductivity curves are presented in fig. 16. As can be seen from the figure (16A), the new pF-curves are steeper than those shown in figure 14A and reflect more of a coarser material (coarse sand).

Simulation was performed using the new pF-curves. The variations of the soil moisture contents with time and depth are presented in fig. 21A - D. The measured values are indicated by the triangular points. As can be observed, the agreement between measured and simulated moisture contents is quite good. It should also be noted that the deeper levels show more deviations between measured and simulated values because of the lateral flow caused by the shallow depth of the infiltration ring and the increasing dominance of finer material.

Fig. 21A - D Variation of measured (points) and simulated (lines) moisture contents with time and depth, plot A

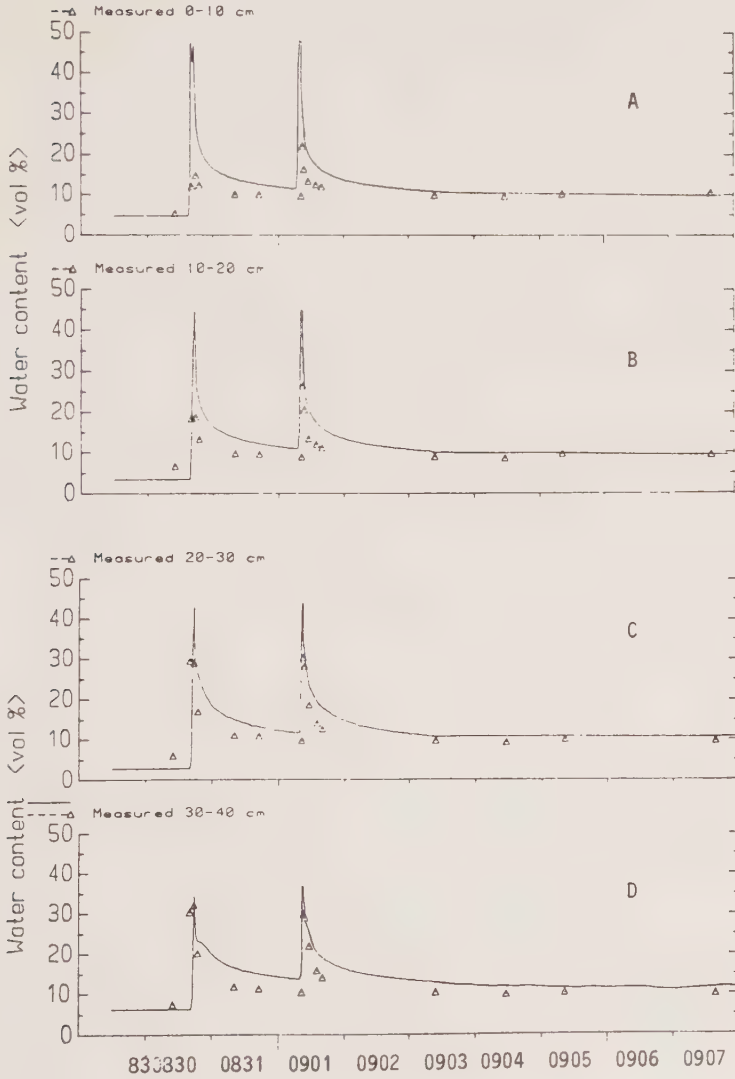
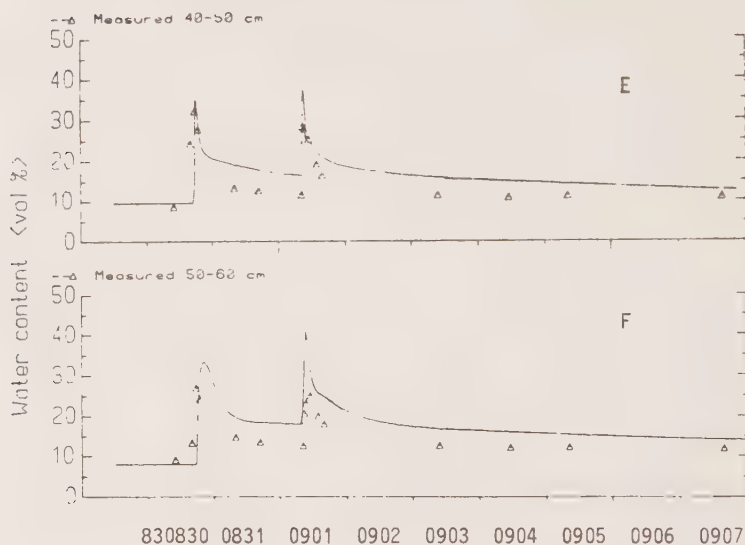


Fig. 21E - F Variations of measured (points) and simulated (lines) moisture contents with time and depth, plot A



10 LONG TERM SIMULATIONS AND RECHARGE ESTIMATES

The idea behind the long term simulations is to compare recharge estimates by the tracer and the water balance methods with those obtained by a mathematical model. The main assumption being that if the soil physical properties are kept constant, then, the variation of recharge with time is mainly dependent on the meteorological parameters such as precipitation and evapotranspiration.

10.1 Simulations

The location of the selected sites is shown in fig. 2. For Uppsala site, the pF-curves (fig. 12A) were used to estimate the necessary soil physical properties. Using meteorological data from Ultuna (2 - 3 km from the study site) and the estimated soil physical properties, soil moisture contents were simulated and the results were compared with those obtained by the gravimetric (only one occasion), the neutron probe and the TDR methods.

Simulation was first performed for the period 1982/83 (period 1) for which observed data were available. The moisture movement and recharge processes for this area and for 1982 have been studied by Saxena and Dressie (1983) using tritium and O-18 as tracers. The recharge estimates by these two tracers are compared with the simulated soil water flows passing the same level as to which the tracers were applied.

The pF-curves for Skediga site are constructed in an indirect way using the soil grain size distribution data. Moisture characteristic curves (pF-curves) for several types of Swedish soils ranging from sand to heavy clays (85% clay content) have been presented by Andersson and Wiklert (1972). By matching the different grain size distributions of the different layers of the study plot with the soil types as presented by these authors, the corresponding pF-curves were selected to describe the field plot. The pF-curves for plot B are shown in fig. 17A. The corresponding estimated hydraulic conductivities are presented in fig. 17B.

In the same way as for Uppsala site, using the given pF-curves (fig. 17A) and the meteorological data from F16 (an air force station located about 1.5 km from the experimental site), soil moisture contents and soil water flows were simulated for the period 1976 - 79 (period 1) for which soil moisture and recharge data are available.

Studies of soil moisture contents using the neutron probe and recharge estimates using the tritium-tagging and the water balance methods for the years 1976 - 79 have been done by the author (Dressie, 1984). In this present study, the simulated moisture contents and recharge estimates for the mentioned period are compared with those obtained by the other methods. The simulation is finally extrapolated to a longer period of 20 years (1961 - 84). A statistical study was performed on the long term simulation results in order to obtain a reliable long term average yearly recharges for the study area.

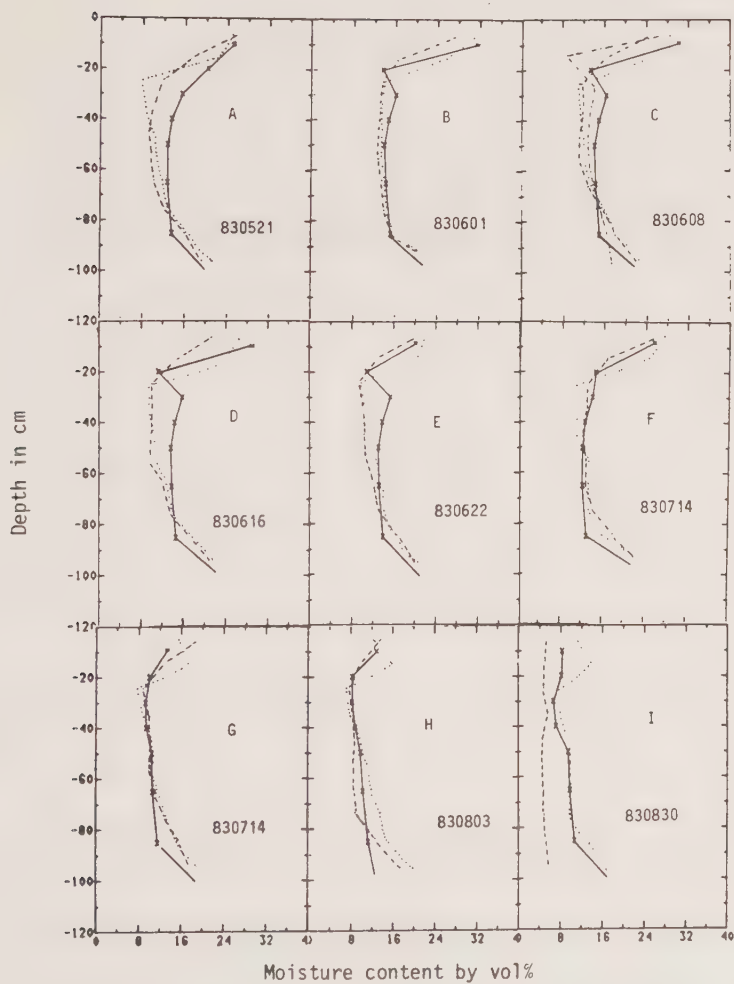
10.2 Results: Plot A - Period 1

The results of the different methods of soil moisture measurements are shown in fig. 22A - I. As shown in the figure the agreement between the different methods is quite satisfactory except during wet and very dry conditions. The curves in fig. 23 show the precipitation and the simulated soil water contents in the profile being considered.

The soil moisture content variations are in direct response to the meteorological conditions. During December and January the soil was frozen at the surface and the unfrozen soil moisture contents as shown in fig. 23B were quite low. The maximum soil moisture contents between March and April corresponded to the snow-melt water. The peak values between June and August, October and November were the responses to the precipitation input of the corresponding periods. These peaks were very much damped by the evapotranspiration losses and were lower than the peak after the snow-melting.

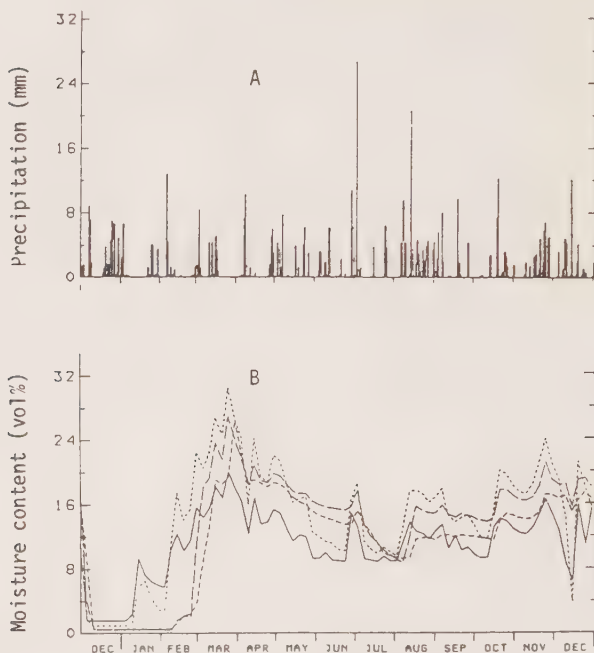
The recharges estimated by the different methods mentioned above are

Figs. 22A - I Simulated and observed moisture contents for the period 830521 - 830830, plot A (Uppsala site).



Legend: = Neutron method ---- = TDR method -.-.- = Gravimetric method — = Simulated

Figs. 23A - B Precipitation and simulated soil moisture content variations for the year 1982, plot A (Uppsala site).



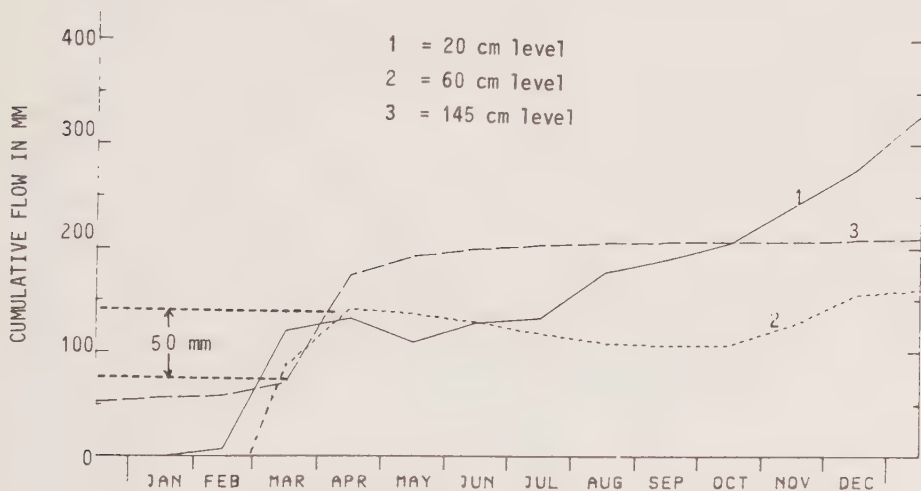
Legend: — = Upper box, ... = 30 cm layer, — — = 40 cm layer, --- = 50 cm layer.

Table 3 Comparison of simulated and measured recharge estimates for plot A and for the period 82 02 12 to 82 05 04.

	Simulated	³ H method	O-18 method
Recharge	50 mm	51 mm	42 mm

shown in table 3. As presented in the table, the simulated estimate of recharge is 50 mm and is comparable to the estimates by the tracers tritium and O-18 (Saxena and Dressie, 1983) which are 51 mm for tritium and 42 mm for O-18. As can be observed, the simulated estimate of recharge is in a very good agreement with the tritium one. The lower value for O-18 method can be due to the difficulty of finding the peak point on the O-18 profile. It has been observed that O-18 profiles do not show the same sharp peaks as the tritium profiles do. This introduces a 5 - 10 cm error in the determination of O-18 peak. In general, considering all the errors involved in the different methods, the agreement is quite good.

Fig. 24 Cumulative flows through the levels indicated and for 1982, plot A (Uppsala).



Simulated cumulative flows at different depths (fig.24) revealed interesting features about recharge conditions during 1982. The three depths indicate a direct response to snow melt-water with a certain time lag between the different levels. At 145 cm depth the increase after about the end of May is small which means that recharge during the summer and autumn was

minor. The decreasing part of curve 2 indicates that most of the moisture at the surface was lost by evapotranspiration and did not contribute to downward percolation.

10.3 Results: Plot A - Period 2

Average yearly recharges for the different regions of Sweden have been estimated by other workers using the water balance method (Melin, 1954 and Tamm, 1959). In order to obtain comparable average yearly recharge and study the seasonal variations, simulation was extended over a longer period. Fig. 26 shows the soil moisture content variations with time and depth which are as expected. The maximum values consistently occur just after the snow melt.

Fig. 26 Soil moisture content variations with time and depth, Uppsala site

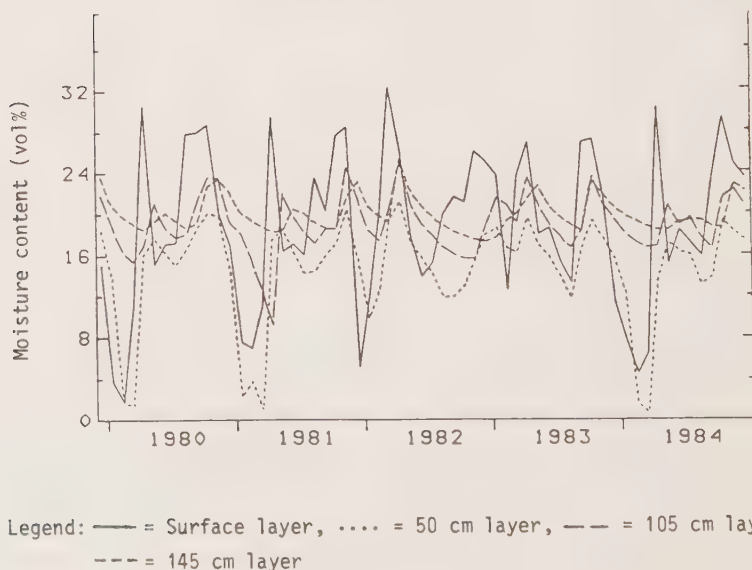
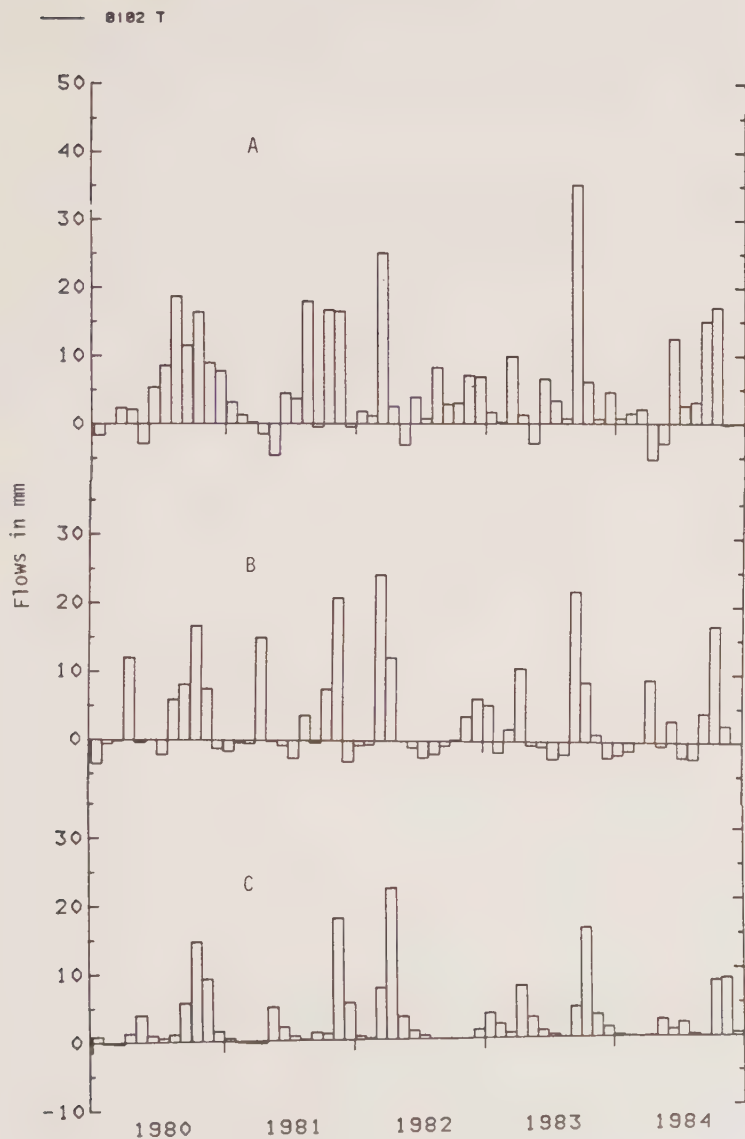


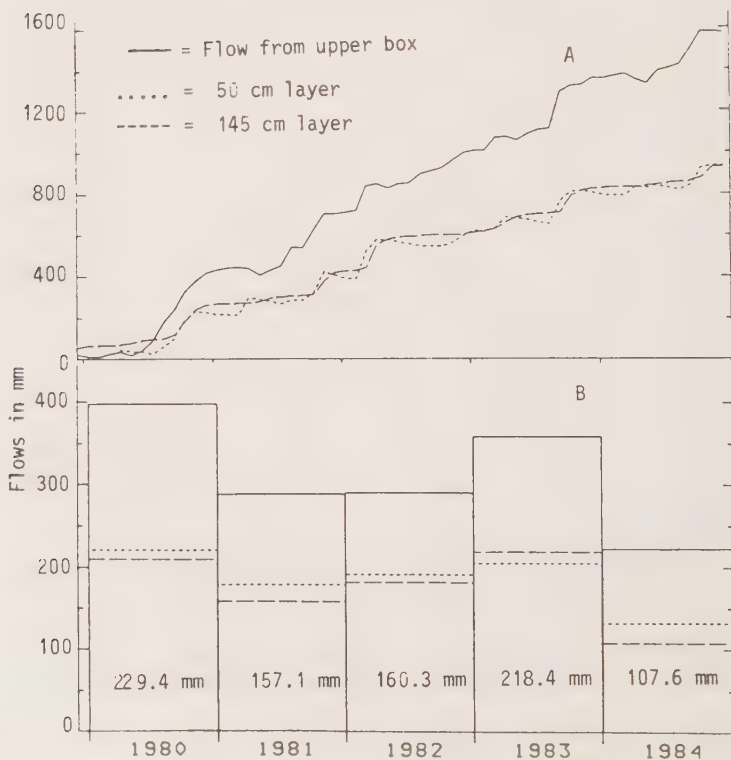
Fig. 27 Monthly sums of flow through the surface (A), 50 cm (B) and 145 cm (C) levels respectively.



The monthly sums of flow through the surface, the 50 cm and the 145 cm levels are shown in figs. 27A - C. It is seen that flow to the deeper layer occurred during late summer and autumn periods in 1980, 1981 1983 and 1984 but, in agreement with the results for period 1, the 1982 recharge occurred only after the snow melt and very little during the rest of the year.

Cumulative percolation rates are presented in fig. 28A and B. The numbers in fig. 28B are the yearly sums of flows passing the 145 cm level and are assumed to be recharge estimates.

Fig. 28 Cumulative flows (A) and yearly sums (B) of flows through the different soil layers.



10.4 Results: Plot B - Period 1

The precipitation (monthly) and evapotranspiration data for the sub-period is given in figs. 29A and B. Illustrated in fig.s 30 and 31 are the simulated and the observed soil moisture contents and their variations with time and depth.

The curves to the left represent the moisture contents as measured by a neutron probe and those to the right are simulated ones. Considering the errors involved in the two methods, the similarity between the observed and simulated values is quite good. Errors involved in selecting the necessary pF-curves to describe the profile and the difficulty in constructing accurate calibration curves for the neutron probe method are examples of such errors.

The temporal variations of soil moisture contents at selected soil depths, as shown in figs. 32A and B, show that observed and simulated values are in good agreement. In the figures the points marked "x" show observed values while the curves represent simulated moisture contents. Simulated values are lower than the observed ones for the winter period, and are caused by different considerations with respect to frozen soils. The neutron probe measurements reflect the total soil water contents whereas the water contents shown in the figure only represent the unfrozen fraction.

Recharge estimates by the tritium-tagging and the water balance methods (Dressie, 1984) are compared with the simulated values as shown in table 5. Column 6 shows average values for the three methods and columns 7, 8 and 9 the deviations of the respective recharge estimates from the averages. As seen from the table, the recharge values by the different methods vary much from each other. The high variations in the 1978 values indicate the failure of the methods for very short periods. Variations of recharge estimates of as high as 35% in the last three columns is in the same range of variations between different methods as found by Losjö (1980). In a comparison of different recharge estimates by different authors, Losjö found variations ranging from 2% to 34%. The high autumn recharges are due to the high autumn precipitations observed.

Fig. 29 Five-day mean air temperatures (A) and monthly sums of precipitation (B) for plot B (Skediga, 1976 - 80).

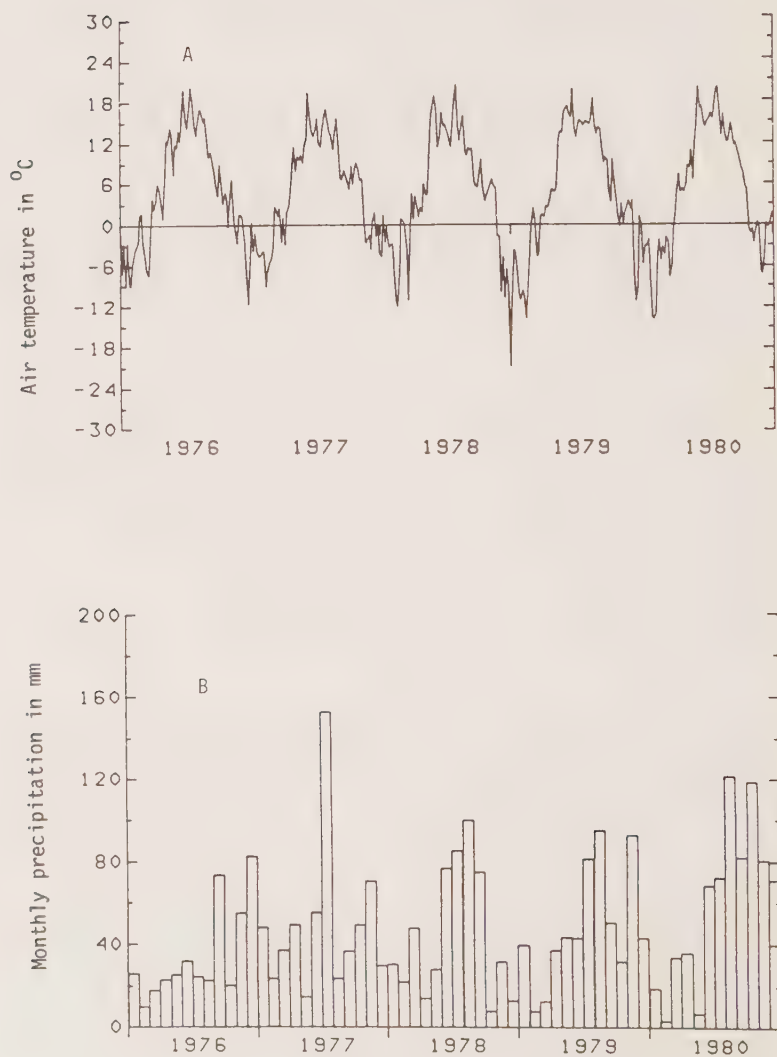


Fig. 30 Observed and simulated soil moisture contents for Skediga plot for the periods 197604 - 7612 (A) and 197702 - 7707 (B).

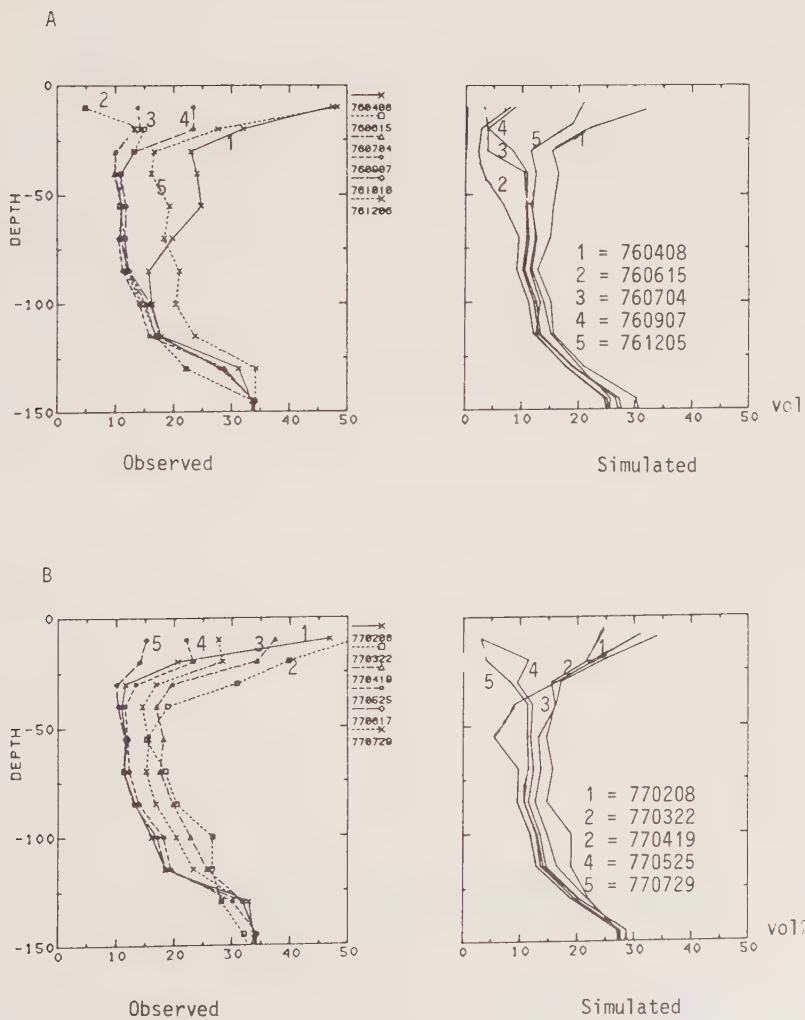


Fig. 31 Observed and simulated soil moisture contents for Skediga plot for the periods 7706-7711 (A) and 7805-7811 (B).

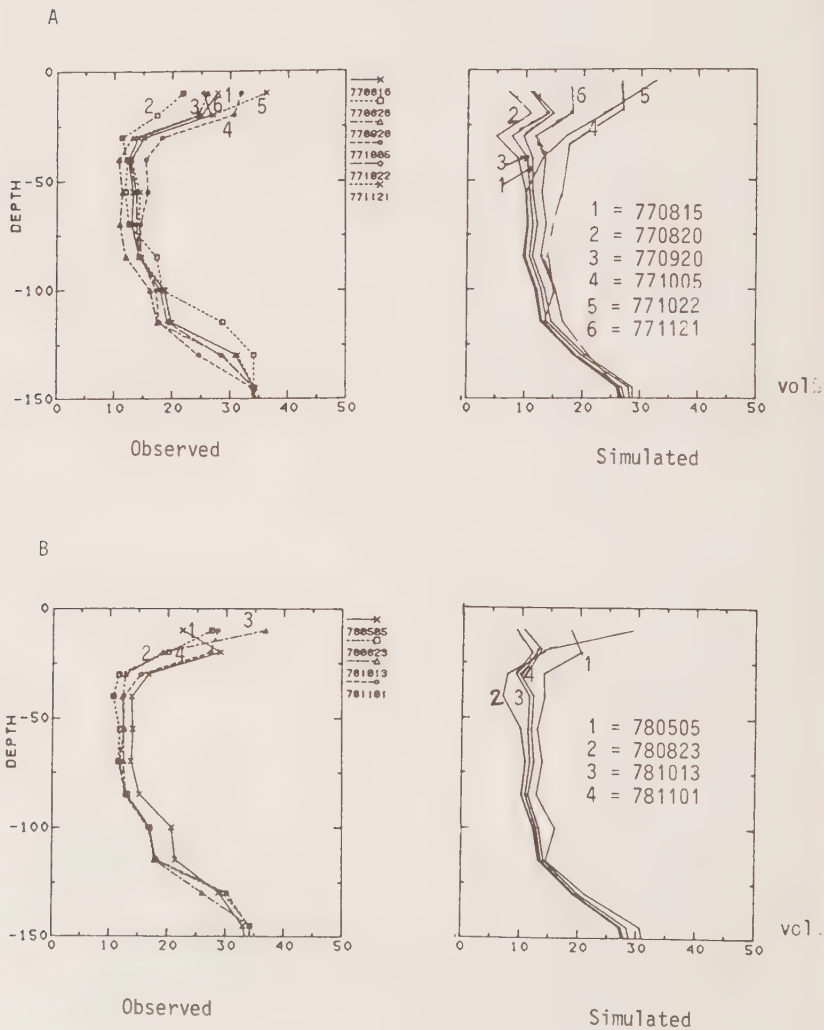


Fig. 32 Temporal variations of soil moisture contents at the 30 cm (A) and 50 cm (B) depths for plot B (Skediga site)

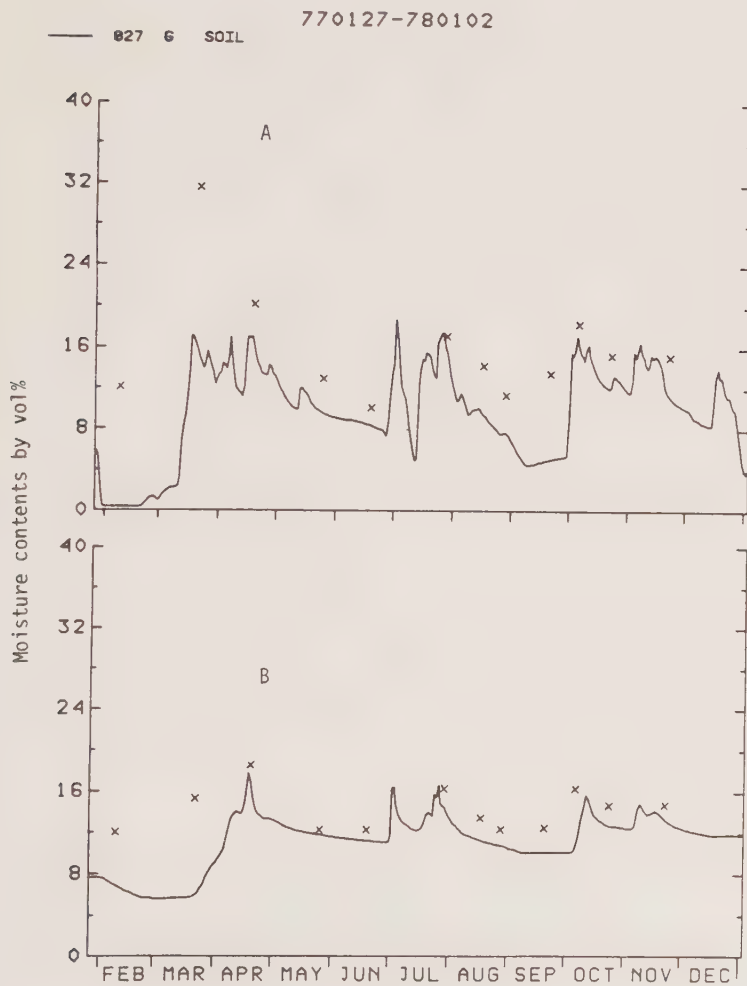


Table 5 Comparison of recharge estimates by the tritium-tagging (^3H), the water balance (Wb) and the simulation methods (Simu) and their deviations from the averages of the three methods.

Period	Days	^3H - method	Wb- method	Simu- method	Ave- rage	Deviations in %			
1	2	3	4	5	6	7	8	9	
760606 - 1003	119	125.2	89.7	63.4	92.8	35	-3	-32	
770630 - 0829	61	129.6	110.8	96.8	112.4	15	-1	-14	
780509 - 0606	28	56.8	26.3	43.9	42.3	34	-35	4	
790903 - 1204	93	138.8	162.8	158.6	153.4	-10	6	3	

10.5 Results: Plot 2 - Period 2

In order to get statistically reliable yearly average recharge estimates and seasonal distributions for the area, long term simulations (23 years) were performed using the same soil physical properties as earlier. The longer the simulation period is the less dependent is the average yearly recharge on the soil physical properties and more dependent on the meteorological conditions (ie recharge = $P - E$). As an input into the simulation model, the precipitation and the evapotranspiration values as calculated by the Penman formula for the whole period are presented in figs. 33 and 34. The figures enable one to distinguish between dry years (eg. 1972) and wet years (eg. 1980).

The simulated temporal variations of the soil moisture content at selected levels are shown in figs. 35 and 36. The pattern of moisture content variations is as expected being very high near the surface and very much damped at the lower depths. It is also observed that the peaks in moisture content consistently occur just after the snow melt.

Fig. 33 Potential evapotranspiration and precipitation values for the given periods, Skediga area.

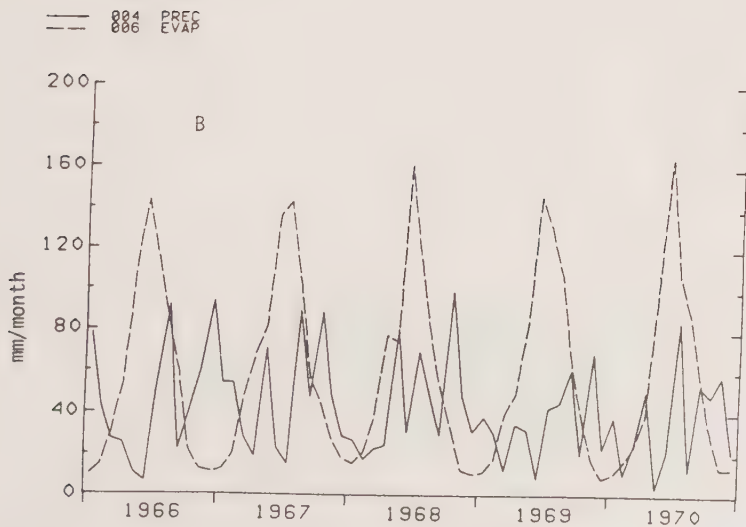
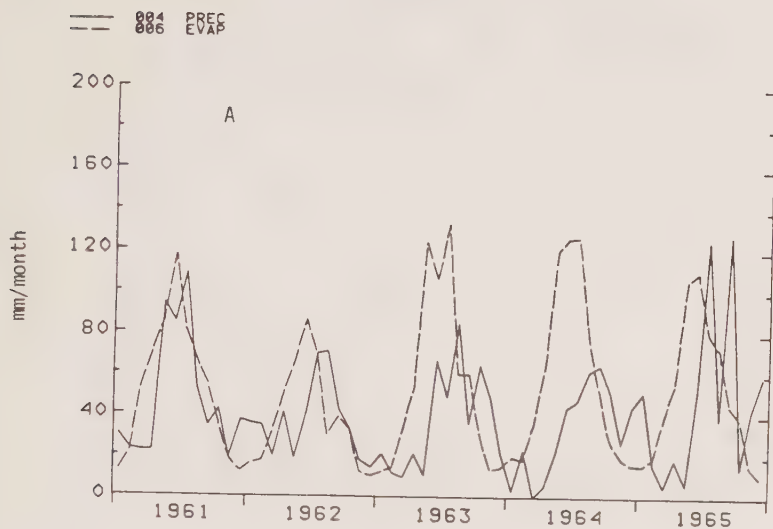


Fig. 34 Potential evapotranspiration and precipitation values for the given periods, Skediga site.

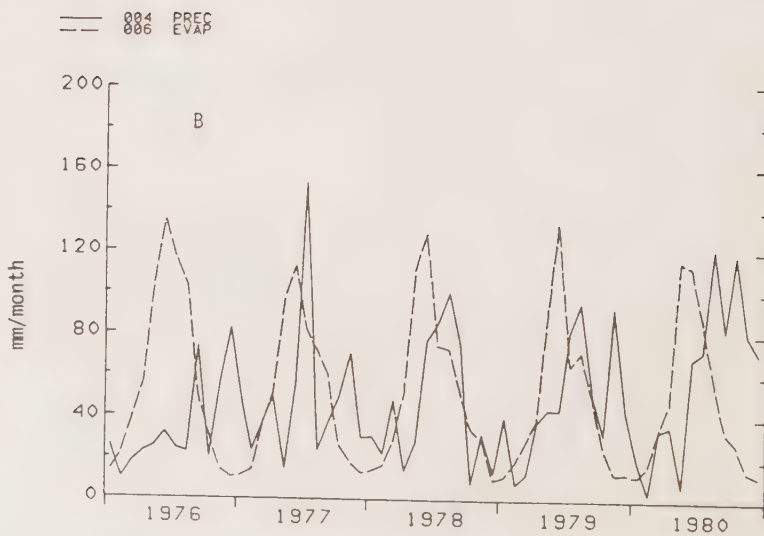
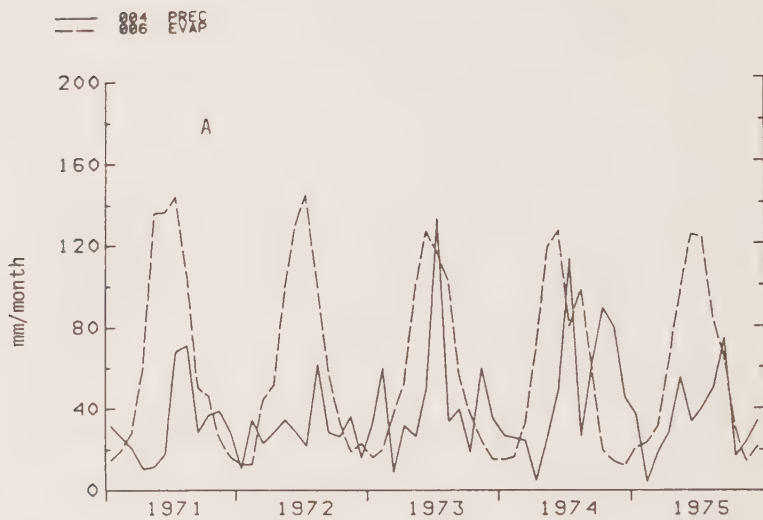


Fig. 35 Soil moisture changes with time and depth (Skediga)

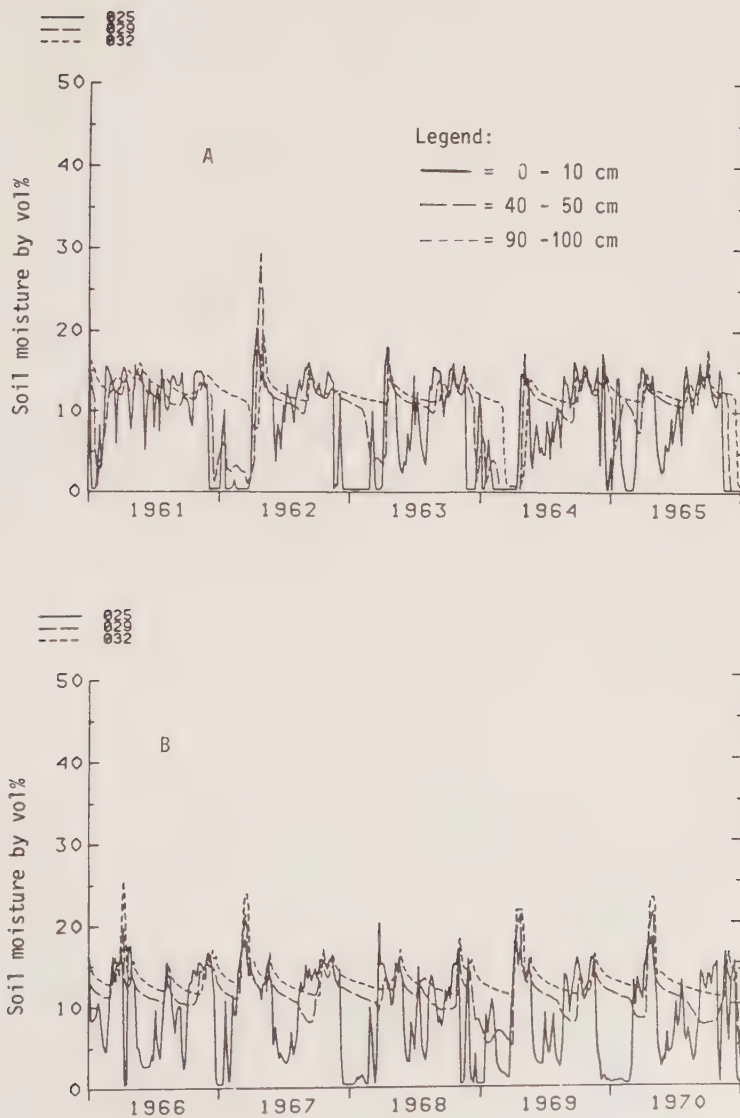


Fig. 36 Soil moisture variations with time and depth, plot A (Skediga site).

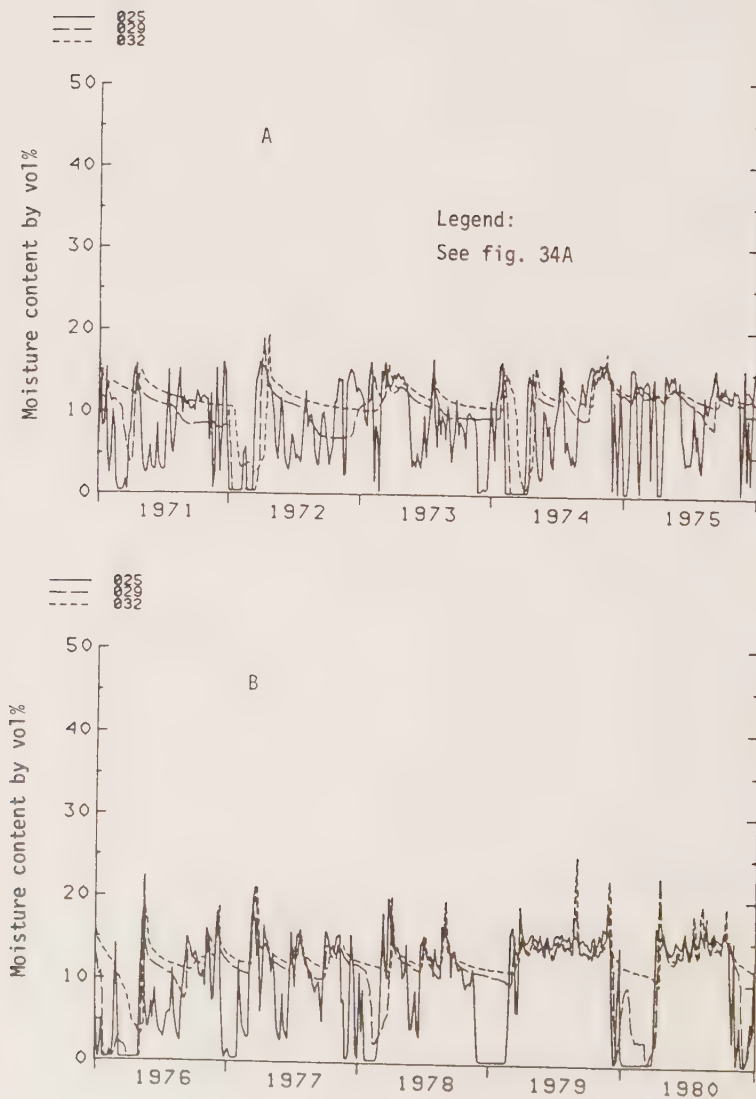
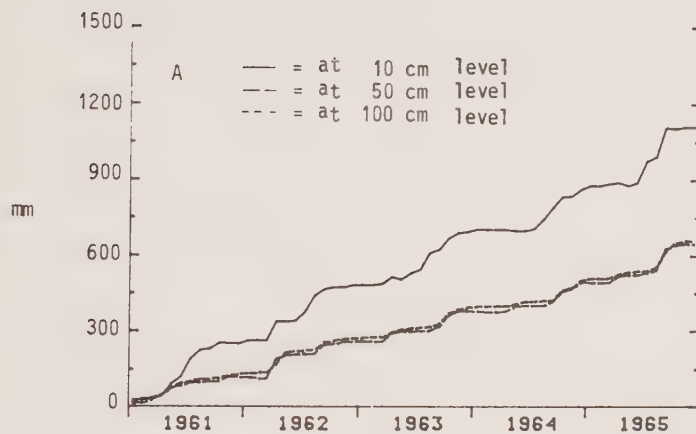


Fig. 37 Cumulative soil moisture flows through the given depths and periods, plot B (Skediga site)



651231-710102

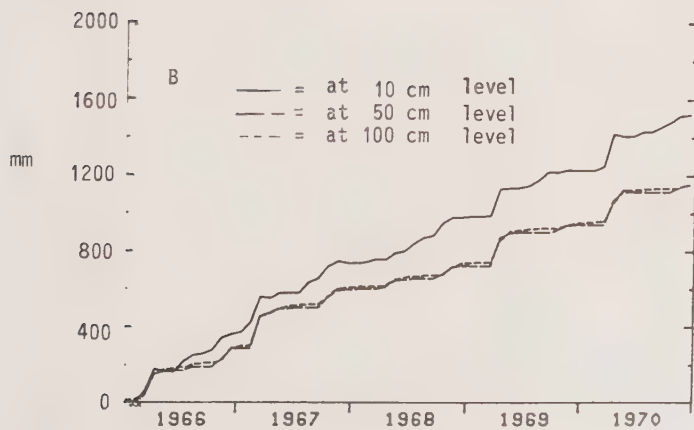
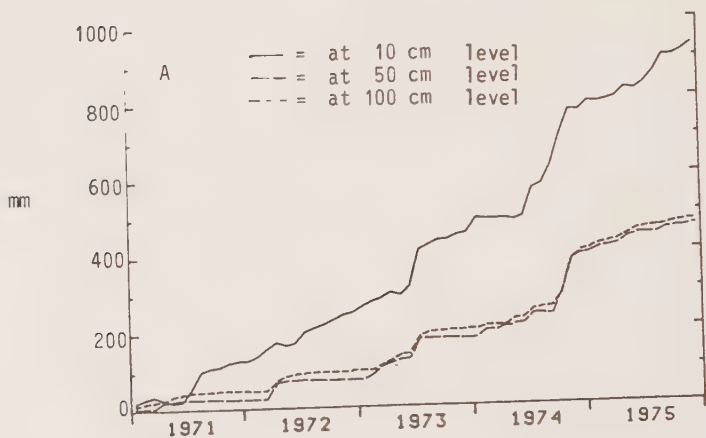
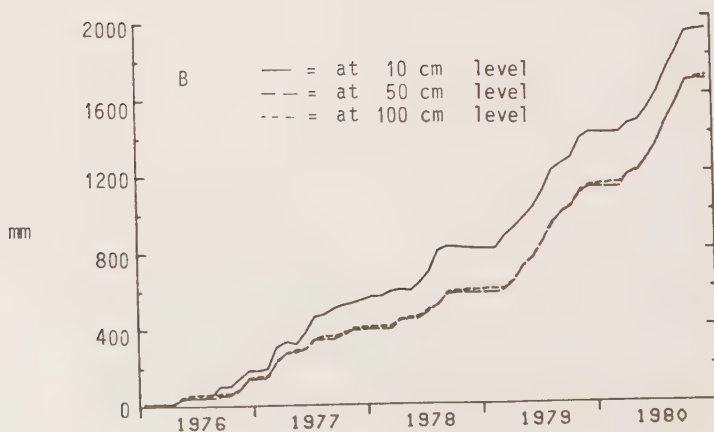


Fig. 38 Cumulative soil moisture flows through the given depths and periods, plotB (Skediga site).



751231-810102



The monthly cumulative sums of soil water flows in the soil profile and at the given depths for the whole study period are as shown in figures 36 and 37. From the figures one can observe that the fluxes through the 50 and 100 cm depths coincide in value throughout the whole period. This proves that moisture below 50 cm depth does not contribute to water-uptake by roots. This was also the assumption when the level for tritium injection was chosen.

10.6 Discussions

10.6.1 Uppsala site

The recharge estimates are summarized in table 4. Comparison of the 50 cm and the 145 cm levels shows that much water is lost by root uptake. The flow through the 145 cm level is assumed to contribute to recharge. In column 6 of table 4, the ratio of recharge to precipitation in per cent is given for each year. Note that the data for 1984 is valid up to the end of October only. As seen from the table, the recharge estimates lie between 20 and 35% of the precipitation input, the higher value occurring in 1982. The higher ratio for 1982 is attributed to the snow cover which was left from 1981. This could have been avoided by taking the hydrological year instead. This effect of delayed snow cover was also observed in the annual recharge estimate for 1982/83 by the 0-18 method which was as high as 44% (Saxena and Dressie, 1983)

Table 4 Simulated flows and recharge estimates for Uppsala area

1	2	3	4	5	6
Year	Flow from	Flow from	Flow from		R
Year	upper box	50 cm box	145 cm box	Prec.	(4/5)
					%
1980	418.8	236.4	229.4	633.9	30.6
1981	288.0	179.2	157.9	641.5	24.5
1982	289.4	191.3	160.3	452.5	35.4
1983	358.2	205.0	218.4	686.2	31.8
1984	222.4	121.9	107.6	502.6	21.4
Average	338.6	203.0	191.5	603.4	31.7

The lower recharge values for 1982 (157.9 mm) and in 1984 (107.6 mm) are also indications of the relatively dry conditions during those years (figs. 25A and B). The 1984 value is the lowest because the simulation was done up to October only. The average yearly recharge of 192.0 mm (excluding 1984) is comparable with SMHI's (1979) and Losjö's (1980) estimates of 200 mm and 193 mm respectively for the same region.

10.6.2 Skediga site

The comparison between yearly precipitation and flow variations during the observation period showed that during very cold winters when the soil is frozen, no flow occurs through the upper 10 cm level until late in March (fig. 39). If a sudden snow melt occurs when the top soil is frozen, then there is a high probability for surface runoff. During mild winters, there occurs flux through the upper layer and contributes to high winter recharges (fig. 40).

The seasonal recharge variations and the corresponding precipitation input are presented in table 6. The annual recharge estimates (T_y), precipitation (P_y) and evapotranspiration ($P_y - T_y$) are also included in the table. The average yearly recharge of 161.7 mm and the average yearly precipitation and evapotranspiration of 514.0 mm and 352.1 mm respectively are compared with values determined by other workers in table 7. As can be seen from the table, the deviations of simulated recharge estimates from those by other workers are in the same range as the variations between those estimated by the different workers. The evapotranspiration value is lower than those by the other methods depending on the soil type (coarse esker material). The areal average evapotranspiration of 391.0 mm falls in the range of values as given by Melin (1954) using the water balance method (table 7). The latest recharge study done by Katarina Losjö (1980) for the same area shows an average value of 193.0 mm and differs from that determined by the simulation method (177.0 mm) by only 9%. Even SMHI's (The Swedish Meteorological and Hydrological Institute) recharge estimate (1979) of 200.0 mm is very near the estimate by the simulation method, differing by only 12%.

Fig. 39 Precipitation (A) and flows through the upper box (G1) and the 100 cm level (G2)

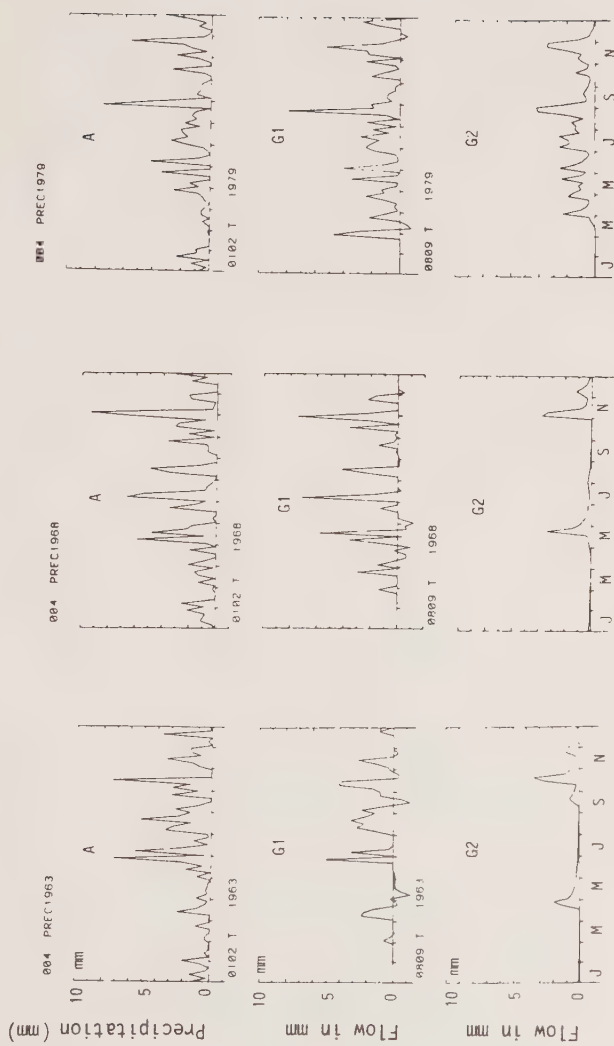


Fig. 40 Precipitation (A) and flows through the Upper box (G1) and the 100 cm level (G2)

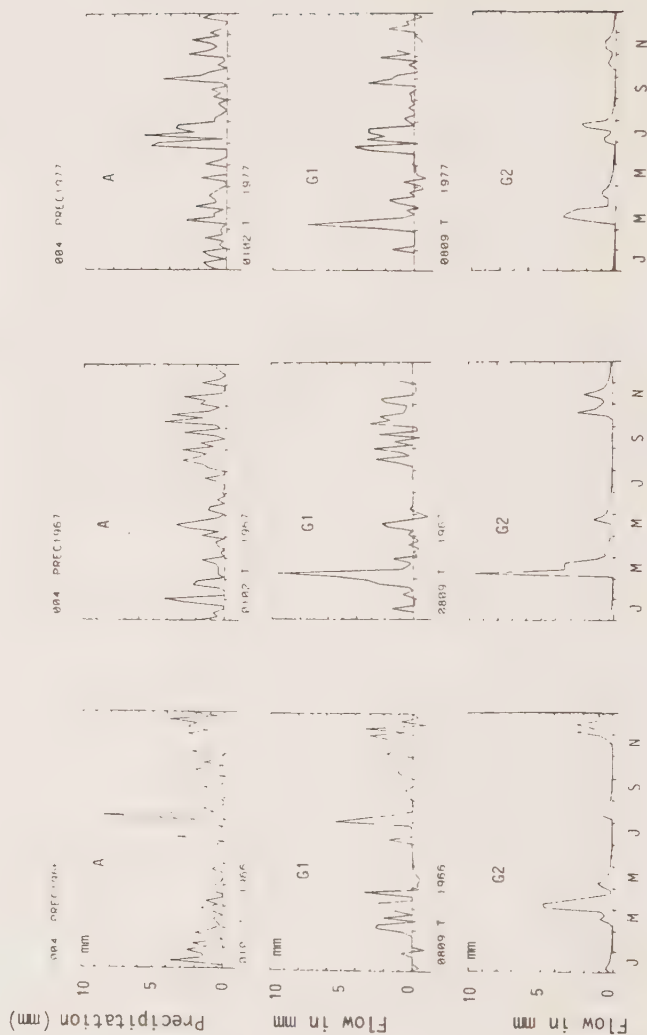


TABLE 6 . SEASONAL RECHARGE VARIATIONS FOR THE PERIOD 1961 - 1983 FOR SKEDIGA AREA. ALL UNITS ARE IN MM

NO Year	Winter 1117 - 0215			Spring 0216 - 0516			Summer 0517 - 0816			Autumn 0817 - 1116			P_y	T_y	$P_y - T_y$ =ET	PET	ET/ PET
	T	P	R	T	P	R	T	P	R	T	P	R					
1. 1961	8.5	75.4	0.12	65.7	201.0	0.28	32.5	195.7	0.17	27.0	98.9	0.27	571	134.0	437.0	627.0	0.70
2. 1962	14.4	89.6	0.16	91.6	100.2	0.33	16.8	183.1	0.09	18.4	65.6	0.28	438	144.2	293.8	463.0	0.63
3. 1963	5.5	42.2	0.13	43.6	98.0	0.45	16.8	168.9	0.10	47.8	133.7	0.35	443	113.2	330.0	657.0	0.50
4. 1964	4.2	25.9	0.16	36.7	71.0	0.52	15.5	173.3	0.09	44.7	122.1	0.37	393	101.1	291.9	701.0	0.42
5. 1965	10.4	72.0	0.14	50.2	76.0	0.78	28.2	292.4	0.10	48.0	115.6	0.41	557	145.8	411.2	607.0	0.68
6. 1966	21.2	148.3	0.14	32.5	42.5	0.77	48.2	165.9	0.29	66.2	195.0	0.34	551	173.1	377.9	668.0	0.57
7. 1967	23.0	137.2	0.18	60.2	112.2	0.54	27.1	152.1	0.18	47.9	166.8	0.29	561	152.2	402.8	762.0	0.53
8. 1968	6.2	66.9	0.10	64.9	135.9	0.48	16.8	148.6	0.11	46.8	179.6	0.26	531	134.7	396.3	692.0	0.57
9. 1969	14.1	81.5	0.17	56.9	75.9	0.78	23.0	149.9	0.15	21.4	112.2	0.19	419	115.3	303.7	722.0	0.42
10. 1970	12.3	74.4	0.16	67.6	75.9	0.89	27.0	151.3	0.18	26.4	127.8	0.21	429	133.3	295.7	671.0	0.44
11. 1971	14.2	77.3	0.18	20.1	39.8	0.51	24.7	167.4	0.15	11.1	103.8	0.11	389	70.1	318.9	781.0	0.41
12. 1972	12.8	68.2	0.19	37.8	91.4	0.41	14.6	111.4	0.15	11.1	78.2	0.14	349	76.3	272.3	727.0	0.37
13. 1973	15.8	101.4	0.16	46.3	107.8	0.43	43.2	205.4	0.21	11.8	113.3	0.10	528	117.1	410.9	702.0	0.59
14. 1974	10.3	76.0	0.14	30.5	80.5	0.38	25.6	160.4	0.13	122.2	213.2	0.57	570	188.6	381.4	662.0	0.58
15. 1975	16.2	112.8	0.14	56.4	116.6	0.48	7.4	220.2	0.05	10.1	74.8	0.14	415	90.1	324.9	697.0	0.47
16. 1976	6.7	53.3	0.13	54.2	80.7	0.67	21.2	120.2	0.18	31.1	157.3	0.20	412	113.2	298.8	694.0	0.43
17. 1977	20.6	109.0	0.19	50.3	119.2	0.59	42.2	213.3	0.21	30.2	150.2	0.20	591	125.3	465.7	593.0	0.79
18. 1978	15.8	100.5	0.16	51.3	118.8	0.43	52.0	259.9	0.20	10.3	52.8	0.20	532	122.4	409.6	626.0	0.66
19. 1979	2.4	60.3	0.04	74.0	124.6	0.59	163.8	227.8	0.72	97.0	168.3	0.58	581	337.2	243.8	556.0	0.44
20. 1980	3.7	55.9	0.07	73.0	111.2	0.66	199.8	275.6	0.72	122.3	270.3	0.45	713	398.8	314.2	572.0	0.55
21. 1981	12.9	97.0	0.13	45.7	110.5	0.41	165.1	205.6	0.80	151.5	295.5	0.51	708	420.8	287.2	595.0	0.48
22. 1982	3.3	100.2	0.03	48.7	105.2	0.46	12.8	142.9	0.09	39.2	139.1	0.28	488	104.8	383.2	671.0	0.57
23. 1983	16.2	126.7	0.13	59.9	147.7	0.41	105.1	286.6	0.37	20.9	139.0	0.15	650	202.1	447.9	666.0	0.67
Max.	23.0	148.3	0.19	91.6	201.0	0.89	199.8	292.4	0.80	151.5	295.5	0.58	713	420.0	447.0	781.0	0.70
Min.	3.7	25.9	0.03	20.1	39.8	0.28	7.4	111.4	0.05	10.3	65.6	0.11	349	70.1	243.8	463.0	0.37
Ave.	11.8	84.9	0.14	53.0	101.9	0.53	48.3	189.6	0.23	46.2	142.3	0.29	514	161.7	352.1	657.0	0.54
σ_{n-1}	5.9	29.2	0.04	15.8	33.7	0.16	54.7	50.7	0.22	40.0	60.2	0.14	101	95.1	63.1	71.2	0.11

T = Recharge; P = Precipitation; P_y = Yearly prec.; T_y = Yearly recharge; ET = Actual evapotranspiration;
 PET = Potential evapotranspiration; R = T/P

Table 7. Comparison of recharge and evapotranspiration values as obtained by the model and other authors.

A

Sources		Recharge	Deviation, %
Melin (1954)		231 mm	24
Tamm (1959)		250 mm	29
SMHI - vattenföring (1979)		200 mm	12
Katarina Losjö (1980)		171 mm	8
		212 mm 193 mm	
		196 mm	
Model	Skediga	161 mm	0
	Uppsala	192 mm 177 mm	

B

Sources		Evapotranspiration
Melin (1954)		350 - 420 mm
SMHI, Eriksson B. (1980)		450 - 500 mm
Model	Skediga	452 mm 391 mm
	Uppsala	352 mm

Table 8. Correlation coefficients for recharge versus precipitation data.

	Winter	Spring	Summer	Autumn	Annual
Linear regression	0.8	0.5	0.6	0.9	0.8
Power regression	-	-	0.6	-	0.8

11 STATISTICAL ANALYSIS

It is logically true that the higher the precipitation fall is the more recharge occurs because then there is more water to account for the different losses such as interception and evaporation. The effect of evapotranspiration on this relation of recharge to precipitation is also interesting. Hence, in order to study the correlation between seasonal recharge and precipitation, regression analysis was performed. The summer and annual data were in addition tested for power regression relationships. The coefficients for the different regression analysis are presented in table 8. The winter (fig. 41) and autumn (fig. 44) recharges are better correlated to precipitation during the corresponding periods than the spring (fig. 42) and summer (fig. 43) recharges which is objectively found from the correlation coefficients being 0.8 and 0.9 for the winter and autumn data but only 0.5 and 0.6 for the spring and summer data respectively.

The power regression analysis on the summer and annual data give the same correlation coefficients as for the linear regression but the curves are approaching the zero recharge line asymptotically giving recharge even with extremely low precipitation values. This is unrealistic because evapotranspiration loss must be compensated before recharge can occur. The linear regression curves cut the precipitation axis at certain threshold values which are roughly equal to evapotranspiration losses for the corresponding periods and below which no recharge occurs. The summer (fig. 43) and annual (fig. 45) threshold values are 162 and 320 mm respectively and these values approximately correspond to the evapotranspiration losses of 141.3 and 352.1 mm respectively (table 6).

The regression analysis on the annual data gives the same high correlation coefficients as for the winter and autumn data showing clearly the high dependence of recharge on precipitation under long term simulations and indicates that such a relationship can be used to estimate recharge for similar areas if precipitation data is available.

Fig. 41 Linear regression line of winter recharge versus winter precipitation.

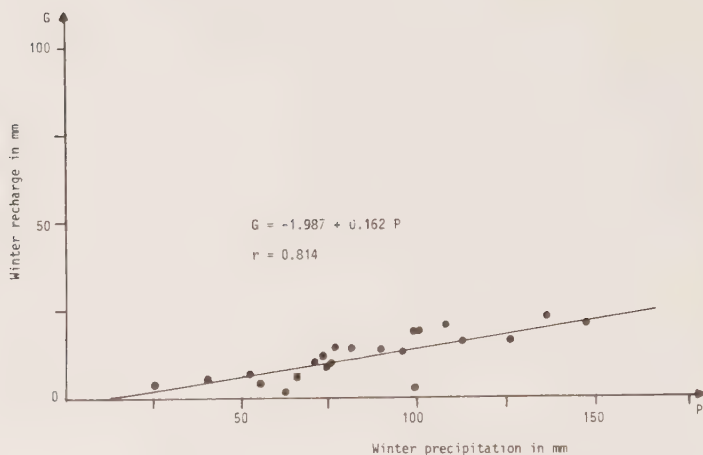


Fig. 42 Linear regression line of spring recharge versus spring precipitation.

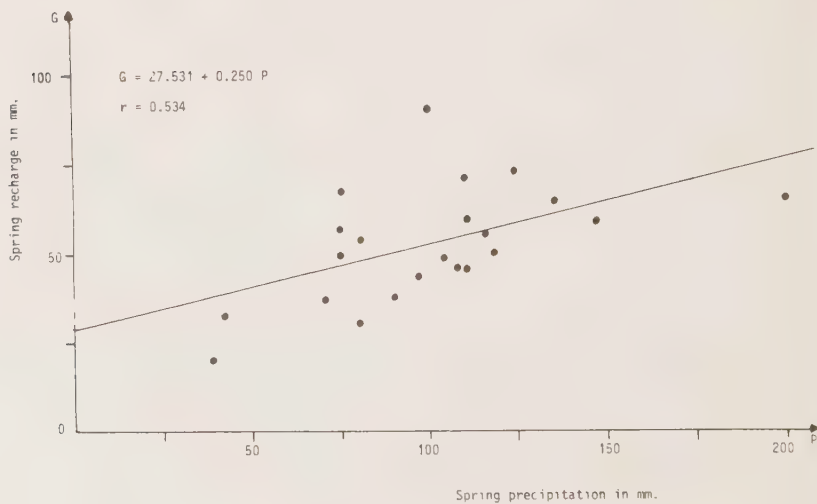


Fig. 43 Linear regression line (1) and power regression curve of summer recharge versus summer precipitation.

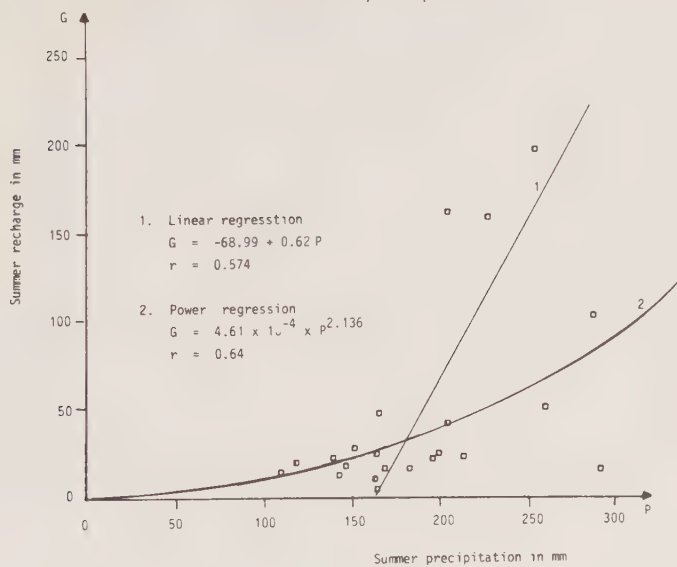


Fig. 44 Linear regression line of autumn recharge versus autumn precipitation.

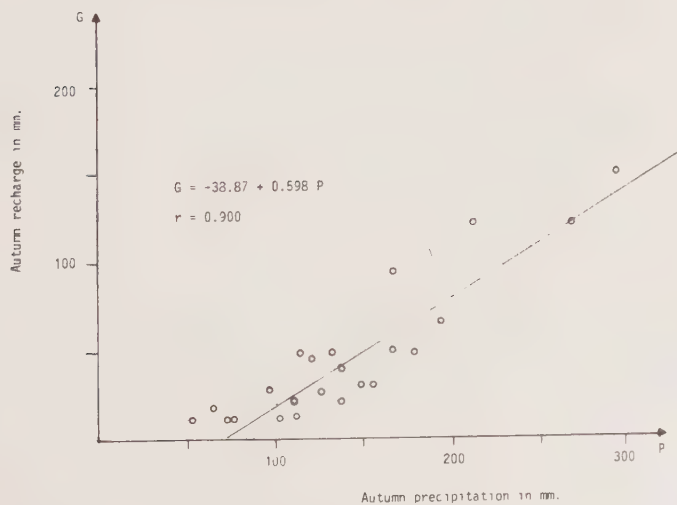
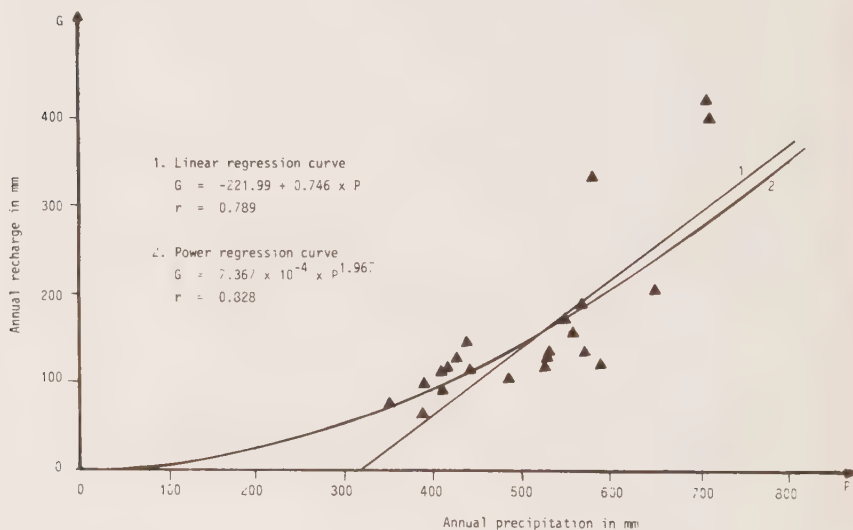


Fig. 45 Linear regression line (1) and power regression curve of annual recharge versus annual precipitation.

Fig. 44 Linear and power regression curves of annual recharge as a function of annual precipitation.



12 CONCLUSION

The infiltration study showed that while a mathematical solution to the unsaturated soil moisture flow equation can predict vertical flows, horizontally homogeneous soil properties on which the solution was based did not completely represent the actual field plot. "Fingering" effect, caused by vertical heterogeneity of the soil and other phenomena (for example faster flows through big pores, cracks and root channels), takes place. These types of flow paths are not accounted for in the model and such phenomena could be found in the deeper wetting front in the observed soil moisture profiles as compared to the simulated moisture profiles. For a further test on infiltration and drainage processes, it should be preferable to determine the soil physical properties with independent methods. Especially the unsaturated conductivity function should be measured. This can be done in the field by using the instantaneous profile method or the crust test procedure.

During a longer period with lower flow rates in the same profile, a good agreement between measured and simulated water contents was observed (fig. 22A - I). The soil moisture content was measured by the TDR, the neutron probe and the gravimetric methods.

The soil water flows and recharge estimates obtained by the long term simulations were compared with recharge values obtained by the tritium and the water balance methods for Skediga area (Dressie, 1984) and by tritium and 0-18 methods for Uppsala area (Saxena and Dressie, 1983). These are presented in tables 3 and 4. The differences in the recharge estimates fall in the same range as observed by other workers.

The long term simulations were made by using constant soil physical properties and variable meteorological data during the study period. It was observed that the variations of the recharge estimates for the

different years are found to be in direct response to the meteorological conditions especially precipitation and evapotranspiration losses.

The regression analysis shows that with long term simulations, recharge is more dependent on the meteorological conditions than on the soil physical properties. One can conclude that a linear relationship can be used to estimate recharge for similar areas if the meteorological data are given. Large differences between areas could however be expected due to differences in biological and micrometeorological conditions.

All the methods compared and used to estimate recharge have their weaknesses. The tritium method is based on the assumption that "piston flow" takes place and that tritium follows the moving water without being detained by the soil material. But in esker formations the vertical flow is usually hampered by the occurrence of clay lenses and texturally different layerings and the presence of clay fractions in the profile can result with a possible tritium interaction with the clay minerals as water diffuses. However, the effect of the latter on recharge measurements is negligible. Dispersion/diffusion processes flatten the tritium peaks whose position in the profile is important in the application of the method. The equation describing the water balance method contains the evapotranspiration term which is difficult to predict accurately. Mathematical solutions do not completely describe the natural flow processes in the soil profile. Therefore, the choice of the method to be applied depends upon the purpose of the experiment. It is also clear that numerical models should be checked by any one of the established field methods. The simulation method can also serve as a tool to study the performance of the experimental work. For example, sensitivity analysis can indicate which properties or field conditions are most important for further investigations. Field experiments are more time consuming and expensive but since pertinent parameters are measured values, the results may be more reliable. The simulation method is also good for obtaining fast information about the general flow conditions in several soil types (with the same soil physical properties).

13 REFERENCES

- Andersson, S. and Wiklert, P. 1972 "Grundförbättring" In: Journal of Agricultural Land Improvement, Prof. Gunnar Hallgren, ed., Institute of Soil Sciences, Dept. of Agr. Hydrotechnics, Uppsala, Sweden.
- Averjanov, S. F. 1950 About Permiability of Subsurface Soils in Case of Incomplete Saturation. Eng. Collect., Vol. 7.
- Babalola, O. 1977 "Field Calibration and Use of the Neutron Moiature Meter on Some Nigerian Soils" In: Soil Science, Vol. 126. No. 2, pp. 118 - 124.
- Beven, K. and Germann, P., 1981 "Water Flow in Macropores II. A Combined Flow Model" In: J Soil Sci., Vol. 32 pp. 15 - 29
- Beven, K. and Germann, P., 1982 "Macropores and Water Flow in Soils" In: Water Resources Research, Vol 18, No. 5, pp. 1311 - 1325
- Briggs, L. 1898 "The Movement and Retention of Water in Soils" In: USDA Yearbook, pp. 399 - 404.
- Brooks, R. H. and Corey, A. T. 1964 Hydraulic Properties of Porous Media. Hydrology paper No. 3 Colorado State University, Fort Collins, Colorado, 27 pp.
- Buckingham, E. 1907 "Studies on the Movement of Soil Moisture" In: USDA Bureau Soils Bulletin 38.
- Brühlhart, A., 1969 Jahreszeitliche Veränderung der Wasserbindung und der Wasserbewegung in Waldböden der Schweizerischen Mittellander Mitteilungen der eidgen. Anstalt fürdas forstliche Versuchswesen Vol. 42. pp. 129 - 237.
- Carneiro, C. and De Jong, E. 1985 "in Situ Determination of the Slope of the Clibration Curve of a Neutron Probe Using a Vol-umetric Technique" In: Soil Science, Vol. 139 No. 3 pp. 250 - 254.
- Corey, G. L., Corey, A. T. and Brooks, R. H. 1965 Similitude for Non-Steady Drainage of Partially Saturated Soils. Hydrology Paper, Colorado State Univ. Fort Collins 3: 1 - 39.

- Corey, A. T. 1977 Mechanics of Heterogenous Fluids in Porous Media.
Water Resources Publications, Fort Collins, Colorado.
- Dasberg, S. and Dalton, F. N., 1985 Time Domain Reflectometry Field Measurements of Soil Water Content and Electrical Conductivity. Soil Sci. Soc. Am. J. 49, pp. 293 - 297.
- Davidson Malcolm, R., 1985 "A Symptomic behaviour of Infiltration in Soil Containing Cracks or Holes" In: Water Resources Research Vol. 21, No. 9, pp. 1345 - 1353.
- Davidson Malcolm, R., 1985 "Numerical Calculation of Saturated - Unsaturated Infiltration in a Cracked Soil" In: Water Resources Research Vol. 21, No. 5, pp. 709 - 714.
- Davis, Stanley, N. and DeWiest, Roger, J. M., 1967 Hydrogeology, John Willy & Sons, Inc. pp. 33 - 38.
- Dressie, Z. 1984 Groundwater Recharge to Eskers. Estimation by the Tritium-tagging and the Water Balance Methods; Study of Soil Moisture Movement. Rep. Ser. A No. 2, University of Uppsala, Division of Hydrology.
- Edwards. W. M. R. R., van der Ploeg and Ehlers, W., 1979 "A Numerical Study of the Effects of Noncapillary-Sized Pores Upon Infiltration" In: Soil Sci. Soc. Amm. J. Vol. 43, pp. 851 - 856.
- Eriksson, E. and Johansson, S. 1978 Gotlands Grundvattenresurser. Avd. för Hydrologi, Uppsala Univ.
- Eriksson, B. 1980 Sveriges Vattenbalans. Årsmedelvärden (1931 - 60) av Nederbörd, Avdunstning och Avrinning. SMHI Rapport Nummer RMK 18 och RHO 21.
- Gardner, W. and Widtsoe, J. 1921 "The Movement of Soil Moisture" In: Soil Science, Vol. 11, No. 3, pp. 215 - 232.
- van Genuchten, M. T. 1980 A Closed Form Equation for Predicting the Hydraulic Conductivity of Unsaturated Soils. Soil Sci. Soc. Am. Journal, Vol. 44, No. 5.

- German, P. and Beven, K., 1981 "Water Flow in Soil Macropores" III. A statistical Approach" In: J. Soil Sci., Vol. 32, pp. 31 - 39.
- Hillel, D. and Gardner, W. R. 1970 Measurement of Unsaturated Conductivity and Diffusivity by Infiltration Through Impeding Layer. Soil Sci. 109: 149 - 153.
- Hillel, D., Krentos, V. D. and Slyianou, J. 1972 "Procedure and Test of an Internal Drainage Method for Measuring Soil Hydraulic Characteristics in Situ" In: Soil Sci., Vol. 114, No. 5, pp. 395 - 400.
- Jansson, P. E. and Halldin, S. 1979 "Model of Annual Water and Energy Flow in a Layered Soil" In: Halldin, S. (ed). Comparison of Forest Water and Energy Exchange Models, pp. 145 - 163. Copenhagen: International Society for Ecological Modelling.
- Jansson, P. E. and Halldin, S. 1980 Soil Water and Heat Model. Technical Description. Tech. Rep. 26, Swedish Coniferous Forest Project, Department of Ecology and Environmental Research.
- Jansson, P-E. 1980a SOIL water and heat model, applied to Möhlin forest. -Proc. from an IUFRO workshop in Zurich, Switzerland, 1979.
- Jansson, P. E. 1980b Soil Water and Heat Model. II Field Studies and Applications. Acta Universitatis Uppsaliensis. Abstract of Uppsala Desertation from the Faculty of Science.
- van Keulen, H. and van Beek, C. G. E. M. 1971 "Water Movement in Layered Soils - A Simulation Model" In: Neth. J. Agrica. Sci., Vol. 19, pp. 138 - 153.
- Kirby, J. M. 1985 "A Note on the Use of a Simple Numerical Model for Vertical, Unsaturated Fluid Flow" In: Soil Sci., Vol. 139. No. 5. pp. 462 - 467.
- Laliberte, G. E., Corey, A. T. and Brooks, R. H. 1966 Properties of Unsaturated Porous Media. Canada Agriculture Research Station, Lthbridge, Alberta.
- Losjö, K., 1980 Elegi Över Vattnets Nedträngande i Marken. C-uppsats i Hydrologi. Avd. för Hydrologi. Uppsala

- Malik, R. S., Kumar, S. and Dahiya, S. 1984 "An Approach to Quick Determination of Some Water Transmission Characteristics of Porous Media" In: Soil Science, Vol. 137, No. 6, pp. 395 - 400.
- Melin, R. 1954 Vattenförlingen i Sveriges Floder. SMHI Meddelanden Serie D Nummer 6.
- Mualem, Y. 1976a "A new Model for Predicting the Hydraulic Conductivity of Unsaturated Porous Media" In: Water Resources Research Vol. 12, No. 3, pp. 513 - 522.
- Mualem, Y. 1976b "Hysteretical Models for Prediction of the Hydraulic Conductivity of Unsaturated Porous Media" In: Water Res. Research Vol. 12, No. 6 pp. 1248 - 1254.
- Mualem, Y. 1982 "Prediction of the Soil Boundary Wetting Curve" In: Soil Sci., Vol. 137, No. 6 pp. 379 - 390.
- Mualem, Y. 1984 "A Modified Dependent-Domain Theory of Hysteresis" In: Soil Sci., Vol. 137, No 5 pp. 283 - 291.
- Philip, J. R., 1957d The Theory of Infiltration: 1: The Infiltration Equation and its Solution. Soil Sci. 83, pp.345 - 357.
- Philips. J. R. 1969 "Theory of Infiltration" In: Advances in Hydroscience" In: Ven Te Chow (ed). Academic Press, New York and London, Vol. 5 pp. 263 - 292.
- Richards, L. A. 1931 Capillary Conduction of Liquids in Porous Mediums. - Physics. 1:318 - 333.
- Saxena, R. and Dressie, Z. 1983 "Estimation of Groundwater Recharge and Moisture Movement in Sandy Formations by Tracing Natural O-18 and Injected Tritium Profiles in the Unsaturated Zone" In: Isotope Hydrology (Proc. Symp. Vienna, 1983), IAEA, Vienna, pp. 139 - 150.
- Schoeller, H. 1962 Les eaux souterraines. Masson & Cie., Paris, 642 p.
- Schudel, P. 1982 "The Accuracy of Measurements of Soil Water Content Made With a Neutron Moisture Meter Calibrated Gravimetrically in the Field" In: Journal of Hydrology, 62(1983) 355 - 361.

- Seckel, K. W. 1978 Feasibility Study for Development of a Transient Three-Dimensional Groundwater Flow Made Utilising the Finite Element Method. University of Maryland, College Park, Maryland 20742, Technical Rep. No. 51.
- Stephens, D. B. and Rehfeldt, K. R. 1985 "Evaluation of Closed-Form Analytical Models to Calculate Conductivity in a Fine Sand" In: Soil Sci. Soc. Am. J. 49: 12 - 19.
- Tamm, O. 1959 Studier Över Klimatets Humiditet i Sverige. Kungliga Skoghögskolans Skrifter Nummer 32.
- Topp, G. G., Davis, J. L. and Aman, A. P. 1982 "Electromagnetic Determination of Soil Water Content Using TDR: I. Application to Wetting Fronts and Steep Gradients" In: Soil Sci., Vol. 46 pp. 672 - 678.
- Wang, J. S. Y. and Narasimhan, T. N., 1985 "Hydrologic Mechanism Governing Fluid Flow in a Partially Saturated, Fractured, Porous Medium" In: Water Resources Research Vol. 21, No. 12, pp. 1861 - 1874.
- Wesley, K. S. 1978 Feasibility Study Development of a Transient Three Dimensional Groundwater Flow Model Utilizing the Finite Element Method. Tech. Rep. No. 51, Water Resources Research Center, University of Maryland.
- Wind, G. P. 1972 "A Hydraulic Model for the Simulation of Non-hysteretic Vertical Unsaturated Flow of Moisture in Soils", In: J. of Hydrology, Vol. XV, pp. 227 - 246.

